

BERICHTE
aus dem
INSTITUT FÜR MEERESKUNDE
an der
CHRISTIAN-ALBRECHTS-UNIVERSITÄT · KIEL

Nr. 294

DYNAMO

Dynamics of North Atlantic Models:
Simulation and assimilation with high resolution models

by
the DYNAMO Group:

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DOI 10.3289/IFM-BER-294

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ISSN 0341-8561



This is the final scientific report of DYNAMO,
a research project supported by the European Union
through the program Marine Science and Technology MAST
(contract number: MAS2-CT93-0060).

Copies of this report can be obtained:

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Contents

Summary	1
Acknowledgements	7
1 Introduction	9
2 Configuration of the Numerical Models	13
2.1 Model domain and horizontal grid structure	14
2.2 Vertical coordinates and bathymetry	16
2.3 Model parameterisation	20
2.3.1 Lateral mixing and bottom form drag	20
2.3.2 Diapycnal mixing	22
2.3.3 Mixed layer concepts	23
2.3.4 Initialisation and surface forcing	26
3 Model Experiments and Analysis	29
3.1 Prognostic model experiments	29
3.2 Model analysis, intercomparison and validation strategy	30
3.2.1 Snapshots and time series for the spin-up phase	32
3.2.2 Data sets for intercomparison phase and sensitivity experiments	33
4 Basin-Scale Overview	35
4.1 Thermohaline circulation	35
4.1.1 Meridional overturning	35
4.1.2 Thermohaline transport at 25° N	40
4.1.3 Water mass properties	49
4.1.4 Meridional heat transport	52
4.1.5 Deep circulation	56

4.2	Aspects of the wind-driven circulation	62
4.2.1	Barotropic transport	62
4.2.2	Near-surface circulation	63
4.3	Upper ocean	69
4.3.1	Ocean-atmosphere fluxes	69
4.3.2	Mixed layer	76
5	Overflow and formation of the Deep Western Boundary Current	83
5.1	Observations	84
5.2	Physical processes and their representation by the models	85
5.3	Former model studies	87
5.4	Model results	88
5.4.1	Large scale circulation patterns	88
5.4.2	Overflow processes	101
5.5	Outlook	109
6	Equatorial Dynamics	111
6.1	Upper layer circulation near the western boundary	111
6.1.1	Summer	113
6.1.2	Winter	116
6.2	Equatorial undercurrents	121
6.3	Deep current field	125
6.3.1	Bottom water flow	132
7	The ventilation of the Central and Eastern North Atlantic Ocean	135
7.1	Introduction	135
7.2	Winter Mixed Layer Characteristics	138
7.3	Surface Circulation Patterns	143
7.4	Isopycnic Potential Vorticity	147
7.5	The Vertical Structure of the Azores Current Near 30° W	167
7.6	Summary and Discussion	176
8	Eddy Variability	181
8.1	Introduction	181
8.2	Geographical distribution of eddy variability	183
8.2.1	Near surface EKE in the tropical and western North Atlantic	183

8.2.2	Eddy kinetic energy in the eastern North Atlantic	193
8.2.3	Two general remarks	196
8.2.4	Sea surface height variability	197
8.2.5	Eddy potential energy	200
8.3	Vertical distribution of eddy energy	204
8.3.1	Vertical section at 48°N	204
8.3.2	Vertical section at 55°W	208
8.4	Discussion	211
8.4.1	Eddy generating processes	213
8.4.2	Eddy lengthscale	214
8.4.3	The role of eddies in the mean circulation	215
8.4.4	Concluding remarks	218
9	European slope processes	219
9.1	Introduction	219
9.2	Near-Surface Circulation Patterns	223
9.3	The Poleward Undercurrent	231
9.4	Conclusions	242
10	Assimilation	247
10.1	Introduction	247
10.2	What can we expect assimilation to achieve?	248
10.3	Assimilation methods	249
10.3.1	Principal existing methods	249
10.3.2	Approaches due to Haines	252
10.3.3	Minimum energy approaches	253
10.3.4	Pre-assimilation	256
10.3.5	The 'cascade' method	257
10.4	The DYNAMO assimilation	258
10.4.1	The altimetric data set	258
10.4.2	The method of OSCHLIES and WILLEBRAND	258
10.4.3	Details of assimilation	264
10.4.4	Simultaneous Assimilation of SSH and SST	266
10.5	The effect of the assimilation: intercomparisons	267
10.5.1	Sea surface height	267

10.5.2 Eddy Kinetic Energy	267
10.5.3 Overall North Atlantic comparison	274
10.5.4 Surface heating and freshening	278
10.5.5 Comparison with observed section	279
10.5.6 Comparison with WOCE current meter statistics	288
10.5.7 Spectra	289
10.6 The seasonal cycle in the assimilation	292
10.6.1 Description of seasonal cycle	292
10.7 Planetary waves in the North Atlantic	297
10.7.1 Background	297
10.7.2 Planetary wave observations	298
10.8 Conclusion/Discussion	299
11 Conclusions	305
References	313
List of Figures	332
List of Tables	333

Summary

The objective of the DYNAMO project has been to contribute to an improved understanding of the circulation in the North Atlantic Ocean, and its variability on synoptic and seasonal time scales. In particular, it has investigated which physical mechanisms control the simulated circulation in high-resolution models, and how sensitive model results are to details of the model formulation. Furthermore, we have investigated the degree to which data from satellite altimetry can improve the simulation of the circulation in high-resolution models.

A main activity of the project has been the development of three different high resolution models which are based on alternative numerical formulations for the vertical discretization, using vertical geopotential levels (LEVEL model), coordinates following surfaces of potential density (ISOPYCNIC), and depth-following sigma-coordinates (SIGMA), respectively. While for the ISOPYCNIC and LEVEL models previous experience was available, the SIGMA model has been the first implementation of its kind in high resolution on the basin scale. An intercomparison of these three models has been performed, with an analysis to assess their ability to reproduce the essential elements of the hydrographic structure and velocity field. Further activities included a determination of the sensitivity of the simulated circulation to details of the wind stress forcing and of the formulation of the upper mixed layer. Observed altimeter data have been assimilated into the LEVEL model, in order to obtain an improved estimate of the basin-scale circulation state.

In the following, important results of the project are briefly summarised:

- *Basin-scale circulation*

All three models have been successful in simulating the North Atlantic circulation with a considerable degree of realism, and can in principle be used for applications which require a realistic dynamical description of the oceanic circulation and hydrographic fields, e.g. in connection with climate problems or with the transport of chemical and biological substances. Differences in performance between the three models are in many aspects smaller, in particular at large scales, than those found in comparisons of coarse-

resolution models. This reassuring result indicates a certain amount of convergence between the models at this resolution. Many problems remain, however, and all three models will require continuing development.

- *Thermohaline circulation*

The principal aspects of the large-scale thermohaline circulation, in particular deep western boundary currents, are simulated well by all models. While meridional overturning and heat transport are controlled by water mass transformation processes in high latitudes, the implementation of the boundary condition at the southern boundary has also a significant influence.

- *Overflows*

In contrast to coarse-resolution models, all models succeed in transporting 4 to 6 Sv of dense water from the Nordic Seas across the Greenland-Scotland Ridge, without an artificial deepening of the ridge topography. The models strongly differ, however, in the diapycnic mixing of the overflow waters south of the ridge. Compared with observations both LEVEL and SIGMA mix too strongly, producing a too buoyant NADW, while ISOPYCNIC lacks diapycnic mixing in this regime and produces a too dense NADW. Because of its crucial role in the large-scale circulation, an improved representation of the bottom boundary layer in this regime should be of highest priority in the future development of all models.

- *Western Boundary Currents*

The pathways of the boundary currents such as the Florida Current and the Antilles Current, but also the North Atlantic Current off Flemish Cap, are strongly influenced by details of topography/geometry. The mechanisms are not well understood, and probably related to insufficient resolution.

- *Air-sea interaction*

All three models are very close in sea surface temperature and heat flux fields. Although the atmospheric forcing (ECMWF-analysis from 1986–89) has been applied in a self-consistent way, systematic deviations from the forcing fields occur in some regions, in particular over much of the subtropical gyre. It appears that in this region the atmospheric forcing is inconsistent with upper ocean dynamics.

- *Surface mixed-layer*

The simulated mixed-layer (ML) depth distribution is a fairly robust parameter, and

simulated well over most regions. Large differences between the models in some regions (such as e.g. the Labrador Sea) are mainly caused by differences in the circulation, not by different ML algorithms. For climate applications, convection is the crucial factor determining winter ML-depth and water mass transformation rates. Detailed simulation of the annual cycle which is most relevant for coupling with biological/chemical models requires a more accurate mixed layer algorithm which includes effects of wind mixing.

- *Equatorial currents*

The meridional overturning is connected to the deep boundary currents crossing the equator which in turn was found to be strongly controlled by the local wind forcing. All models also show a deep seasonal response to wind forcing.

- *Thermocline ventilation in eastern North Atlantic*

All three models include the relevant processes for subduction and ventilation of the main thermocline, and perform fairly well in this region, in particular ISOPYCNIC. It appears that the frontal zone associated with the Azores Current has a crucial role in acting as a partial barrier to the southward ventilation of Eastern North Atlantic Water, which forms near the European continental margins.

- *Eddy energy*

For the first time it has been possible to directly compare mesoscale eddy activity in three different models. Due to insufficient resolution, the eddy field cannot be fully resolved. As a consequence, the eddy activity in all models is too low, and strongly depends on subgrid-scale parametrization. The eddy energy is generally correlated with the mean kinetic energy, except on continental slopes where topographic control has a stabilizing influence.

- *European slope processes*

Although not specifically designed for the European slopes, all three models have performed surprisingly well in simulating important aspects of the seasonal cycle of near-surface circulation and temperature fields, and in revealing the northward spreading of saline water of Mediterranean origin. Even higher salinities near the Iceland-Scotland ridge likely result from downward mixing from the surface layers, and do *not* originate from Mediterranean Water.

- *Assimilation of altimeter data*

The assimilation of combined ERS-1 and Topex/Poseidon altimetric sea-surface height *anomaly* data has generally improved the simulation, in particular with respect to the mesoscale eddies and to wave propagation. However, unrealistic patterns in the *mean* circulation and in the water mass distribution could not be corrected substantially. It is concluded that improvements will in addition require assimilation of the *mean* SSH (altimetric or other) and also of in-situ observations of water mass structure.

- *Specific characteristics of the LEVEL model:*

- flexible implementation for a wide range of applications
- lowest CPU-time of all models in present configuration (factor 2 resp. 2.5 to the other models), hence best suited for sensitivity studies and assimilation experiments
- only model with open southern boundary
- excessive unphysical diapycnal mixing in outflow region of Denmark Strait overflow, due to poor representation of flow over bathymetry, and also in regions of strong isopycnal slopes
- meridional overturning and heat transport significantly weaker than in other models, due to mixing in outflow regime and spurious upwelling in midlatitudes
- somewhat unrealistic pathway of North Atlantic Current, very sensitive to topographic details

- *Specific characteristics of the ISOPYCNIC model:*

- isopycnic concept optimal for water-mass spreading along potential density surfaces
 - circulation in several areas more realistic than in other models, e.g. in main thermocline and in North Atlantic Current region
 - no unphysical diapycnic mixing, and strength of NADW meridional overturning and heat transport simulated well, but lower branch too deep and dense due to a lack of diapycnic mixing (entrainment) in the outflow region
 - single potential density dynamically inconsistent, leads to deviations from thermal wind relation and e.g. prevents proper simulation of circulation and water mass distribution associated with the AABW
-

-
- eddy kinetic energy factor of 4 lower than in both other models, probably caused by too much lateral mixing
 - unable to permit velocity shear even in deep mixed layer
 - *Specific characteristics of the SIGMA model:*
 - coordinate concept optimal for topographically dominated flows
 - first application of SPEM (Sigma-coordinate Primitive Equation Model) to basin-wide circulation at eddy resolution, still less tuned than other two models
 - circulation generally more vigorous than in other models, especially near bottom (overflow), and numerous small-scale recirculation gyres in vertically integrated flow
 - strength of overturning cell and meridional heat transport in subtropics simulated well, but sinking/formation of NADW not concentrated in subpolar region
 - unphysical diapycnic mixing, in particular in regions of strong isopycnal or topographic slopes, due to formulation of lateral diffusion along geopotentials
 - some unrealistic aspects related to strong topographic control, such as e.g. deviations from Sverdrup balance in subtropical gyre and flow of warm North Atlantic Current water into Labrador Sea
 - limited to strongly smoothed bathymetry, in order to avoid numerical errors in calculation of pressure gradient
-

Acknowledgements

The DYNAMO project would not have been possible without support from various other programs, and from related research efforts of other groups or persons which have preceeded the project or have been running simultaneously.

The development of the LEVEL-model in Kiel grew out of the activities in the WOCE Community Modeling Effort (CME) which was designed in the international WOCE program and has been supported through the Sonderforschungsbereich 133 and through institutional resources of the Institut für Meereskunde. Part of the development of the algorithm for the assimilation of altimetric data has also been supported by the SFB 133. Hydrographic data for the eastern North Atlantic have been kindly provided by Dr. M. Knoll from the Abteilung Meeresphysik at IfM Kiel. The 48°N WOCE hydrographic section was kindly given to us prior to publication by Dr. P. Koltermann from BSH Hamburg. An analysis of the surface drifter data for the southern part of the DYNAMO domain has been partially supported through the german WOCE project. Spin-up and intercomparison run with the LEVEL-Model were carried out at the Deutsches Klimarechenzentrum, with support from the Bundesministerium für Bildung und Forschung. Assimilation experiment and sensitivity studies were carried out at the Rechenzentrum der Universität Kiel.

The development of the ISOPYCNIC model builds on work undertaken in the AIM ("Atlantic Isopycnic Model") project, part of which was funded by the United Kingdom Ministry of Defence, and part of which was funded through the Science Budget from the United Kingdom Government in the context of WOCE. A large part of the Southampton DYNAMO project was also similarly funded through the Science Budget from the United Kingdom Government in the context of WOCE. The assistance of R. Bleck, E. Chassignet and L. Smith (all at University of Miami), and also of R. Marsh, M. Huddleston and L. West in setting up our earlier models in the AIM project has been very helpful. Scientists at the Hadley Centre for Climate Prediction and Research, part of the United Kingdom Meteorological Office, principally Richard Wood, Chris Gordon, and Malcolm Roberts, participated in an earlier model intercomparison from

which much was learned about the respective models. The assistance and help of staff at the Atlas Centre of the Rutherford Appleton Laboratory in the UK was crucial for implementing the ISOPYCNIC model run on the CRAY-YMP there, and for the subsequent data management. Altogether, the runs used about 10,000 single processor hours which were provided as a national resource without charge to the DYNAMO project.

The DYNAMO activities at the Grenoble group at LEGI have also benefitted from several interactions with the french WOCE program, funded by Programme National d'Etude de la Dynamique du Climat (PNEDC). The atmospheric forcing function used for the project has been developed for both WOCE and DYNAMO objectives, with means from both programs. The development of the SIGMA model has been carried out by a group of several academic institutions within the WOCE community, with major contributions from Rutgers University-New Brunswick, LEGI-Grenoble, AWI-Bremerhaven and CSIRO-Hobart. The application of the sigma-coordinate to the DYNAMO configuration greatly benefitted from progress made within WOCE. The principal scientists in the Grenoble group (Le Provost, Barnier, Molines) are supported by Centre National de la Recherche Scientifique (CNRS). The data provided by the Grenoble group for the assimilation experiment were processed from a merged Sea Level Anomaly data set for both TOPEX/Poseidon and ERS-1 satellites, computed and made available by CNES/CLS (P.-Y. Le Traon, Toulouse). The common graphics software used to produce the intercomparison figures was developed in Grenoble within the program SIMAN (SIMulation de l'Atlantique Nord) funded by DRET (Direction de la Recherche et Etudes Techniques of the french ministry of Défense). Computer resources for the Grenoble group were provided by IDRIS (Institut pour le Développement des Ressources en Informatique Scientifique, CNRS), at a total of 2169 hours of C98 (monoprocessor) computing time over the duration of the project.

We gratefully acknowledge all these sources of support.

Chapter 1

Introduction

Observational programmes in the North Atlantic have successively revealed that the large-scale current systems are dominated by a rich spectrum of variability. The limitations of sea-going data acquisition, even using modern techniques such as ship-based ADCPs, remotely-tracked drifting buoys or long-term moorings, mean that very little of this variability may be sampled directly at sea. Conversely, satellite observations provide essentially global coverage of synoptic data, but only at the top of the ocean. Accordingly, a quantitative determination of the circulation, let alone an understanding of its dynamics and interactions, cannot be obtained through field studies alone; numerical models are necessary both to help interpretation of the data, and to build an understanding of ocean dynamics.

Studies of ocean circulation by means of numerical models have certainly proliferated in recent years, fueled both by the recognition of an increasing realism of the model solutions and by the increases in computing power (see, e.g., MCWILLIAMS, 1996, for a review of the history and solution behaviour of ocean circulation models). However, the vast range of time and space scales excited in the ocean continues to present a formidable challenge to ocean modelling. Though it has become possible in recent years to incorporate a significant fraction of the mesoscale eddy spectrum into models of global coverage (e.g., STAMMER et al., 1996), the integration period of these models has to be limited to a few decades, and the solutions continue to be dependent on parameterisations of important, small-scale physical processes.

A suite of sensitivity studies into the effect of different physical processes and model parameterisations in the context of a high-resolution North Atlantic model has been performed under the US-German “Community Modelling Effort” (CME) in support of the World Ocean Circulation Experiment (for a review see BÖNING and BRYAN, 1996). The basic model configuration was that of an Atlantic basin between 15°S and 65°N, with closed walls at the

northern and southern boundaries. A focus of the CME analyses was on the dynamics of the current variability in the tropical and subtropical Atlantic, where the model solutions began to reproduce many observed features in a realistic way (e.g., DIDDEN and SCHOTT, 1991; SCHOTT and MOLINARI, 1996). There were considerable problems, on the other hand, at higher latitudes. Perhaps the most glaring deficit the CME shared with a number of other model studies concerns the failure to simulate the observed current structure (REDLER and BÖNING, 1997) and eddy variability (TREGUIER, 1991) in the north-eastern parts of the basin. Sensitivity experiments with different versions of open boundary conditions emphasised the need for an improved representation of the water exchange with the Norwegian Sea (REDLER and BÖNING, 1997). The numerical representation of the outflow of deep water across the Greenland-Iceland-Scotland ridge in turn was found to be of decisive influence on important, basin-scale aspects of the circulation; in particular, the density of the outflow prescribed by the northern boundary condition in the CME was a key factor determining the structure and strength of the meridional overturning circulation with its associated northward transport of heat (DÖSCHER et al., 1994; HOLLAND and BRYAN, 1994).

To date, a majority of such simulations have been produced using one standard type of ocean model, i.e., the model developed at GFDL (BRYAN, 1969; COX, 1984; PACANOWSKI, 1995). The dependence of model solutions on small-scale physical processes such as overflows, or on the parameterisation of sub-grid-scale mixing (e.g., BRYAN 1987), emphasises the need to critically assess the potential of alternate model formulations. New model developments particularly include formulations with a different representation of the vertical coordinate. In the isopycnic coordinate system, the thickness of individual (and homogeneous) layers of fluid is predicted rather than the density at fixed level. In the terrain-following (sigma) coordinate system the model domain smoothly conforms to the irregular bottom.

Several systematic investigations of model performance on the basin-scale have been in the form of an intercomparison, where different models or algorithms were run under similar conditions. The important role of the numerical advection algorithm on the basin-scale North Atlantic circulation has been discussed by GERDES et al. (1991). CHASSIGNET et al. (1996), and ROBERTS et al. (1996) have performed an intercomparison between the GFDL-model and the isopycnic model, and found quite substantial differences as a consequence of the different vertical discretisation. All these studies were however at a coarse resolution of $1\text{--}2^\circ$, and it is not clear to what extent the results also hold for high resolution simulations. The DYNAMO project represents the first attempt to systematically compare and test the ability of three high-resolution models with different numerical techniques to reproduce

the essential elements of the North Atlantic circulation.

Significant progress in the understanding of the dynamics of the North Atlantic circulation depends on improvements in model simulations. An important step toward this goal is an identification of the critical model factors determining the numerical solutions in different dynamical regimes, and on different space and time scales. Another approach involves the utilisation of the the great potential of satellite altimetry for observing the ocean circulation. Altimetric measurements are particularly useful for determining aspects of the mesoscale variability as variations in sea surface height are directly related to the surface geostrophic velocity. A combination of altimetric observations with dynamical models by data assimilation methods has been explored mainly in idealised applications, i.e., models with rather limited physics or idealised basins (e.g., VERRON and HOLLAND, 1989; SCHROETER et al., 1992); there has however been little experience with assimilation in high-resolution, primitive equation models of full ocean basins (OSCHLIES and WILLEBRAND, 1996).

The DYNAMO project aimed at an improved simulation of the North Atlantic Ocean, by combining high-resolution prognostic models with data from altimetric observations. It specifically involved an intercomparison of three different high resolution models based on alternative numerical formulations (using vertical levels, isopycnic coordinates, and depth-following sigma-coordinates), by assessing their ability to reproduce the essential elements of the hydrographic structure and velocity field; a determination of the sensitivity of the ocean circulation to the most important forcing function at the synoptic to seasonal time scales, i.e. the wind stress; assimilation of observed altimeter data into a high resolution simulation of the basin-scale circulation to obtain an improved 'state estimation' of the hydrography and velocity field.

Chapter 2

Configuration of the Numerical Models

One of the main objectives of the DYNAMO project is an intercomparison of state-of-the-art primitive equation ocean circulation models as a tool to improve the simulation of the circulation in the North Atlantic Ocean with respect to the variability on synoptic and seasonal timescales, to the response and sensitivity to atmospheric forcing, and the to role of eddies and fronts.

Thus, three models, following different concepts for the discretisation of the vertical coordinate, are setup as close as possible for a systematic model intercomparison:

- a classical cartesian (or z-coordinate) model based on the GFDL-MOM code (COX, 1984) at IfM Kiel,
- an isopycnal coordinate model based on the MICOM code (BLECK and CHASSIGNET, 1994) at Southampton Oceanography Centre,
- a depth following sigma-coordinate model based on the SPEM code (HAIDVOGEL et al., 1991) at LEGI-IMG Grenoble.

The models will henceforth be referred to as LEVEL, ISOPYCNIC and SIGMA, respectively.

LEVEL may be regarded as a successor of the well established Kiel CME model family, which consists of a hierarchy of coarse ($1.2^\circ \times 1^\circ$) and eddy-resolving ($0.4^\circ \times 1/3^\circ$ and $0.2^\circ \times 1/6^\circ$) horizontal resolution configurations, covering the tropical, subtropical and sub-polar North Atlantic. These models have been used to explore the ocean circulation by means of an investigation of the sensitivity of the model solutions to different atmospheric forcing functions, lateral boundary conditions, horizontal resolution, and the parameterisation of subgrid scale processes (e.g. BÖNING et al., 1991, DÖSCHER et al. ,1994, BÖNING and HERRMANN, 1994, BECKMANN et al., 1994). Methods have been developed to assimilate GEOSAT

altimeter data (OSCHLIES and WILLEBRAND, 1996). Furthermore, a regional eddy-resolving model of the subpolar North Atlantic using open boundary conditions has been used for tracer studies and to explore in more detail subpolar dynamics (REDLER and BÖNING, 1997).

ISOPYCNIC benefits from the experience of the Isopycnic Modelling Group at the Southampton Oceanography Centre, and is the successor of the Atlantic Isopycnic Model (AIM). Coarse resolution (1°) versions of AIM have been setup in a domain extending further north than CME to explore the dynamics of overflow processes in a model intercomparison study with the Hadley Centre, using a GFDL-type level model (NEW et al. 1995, WOOD et al., 1996). An eddy-resolving ($1/3^\circ$) model of the North Atlantic has been run before the DYNAMO project begun.

In contrast to the other models, SIGMA has mainly been used in idealized process studies (CHAPMAN and GAWARKIEWICZ, 1995, BECKMANN and HAIDVOGEL, 1997). The application of SPEM in DYNAMO is the first time eddy-resolving, basin scale application of this model, and thus a completely new attempt. Nevertheless, the experience gained at LEGI in coarse resolution ($4/3^\circ$ isotropic grid) simulations of the South Atlantic performed within the WOCE-funded MOCA project (BARNIER et al., 1997, MARCHESIELLO et al., 1997), provides a platform for the model development and adaptation to the North Atlantic configuration.

2.1 Model domain and horizontal grid structure

All models share a horizontal domain extending from approximately 20°S to 70°N , and 100°W to 16°E and thus cover the tropical, subtropical and subpolar North Atlantic Ocean (see figure 2.1).

Both SIGMA and ISOPYCNIC make use of the C-grid, whereas LEVEL is based on a discretisation on the ARAKAWA B-grid. To yield an efficient numerical realisation, the number of grid-points in the zonal and meridional directions is slightly different for the three models. This results in very small and negligible differences in the domains covered by the various models which remain within the limits defined above.

All three DYNAMO models solve the physical system of the primitive equations on an isotropic horizontal grid with a resolution of $1/3^\circ$ at the equator. The isotropy of the grid is realized with a grid-step in longitude, $\Delta\lambda$, which is constant:

$$\Delta\lambda = \Delta_{deg} = 1/3^\circ \quad (2.1)$$

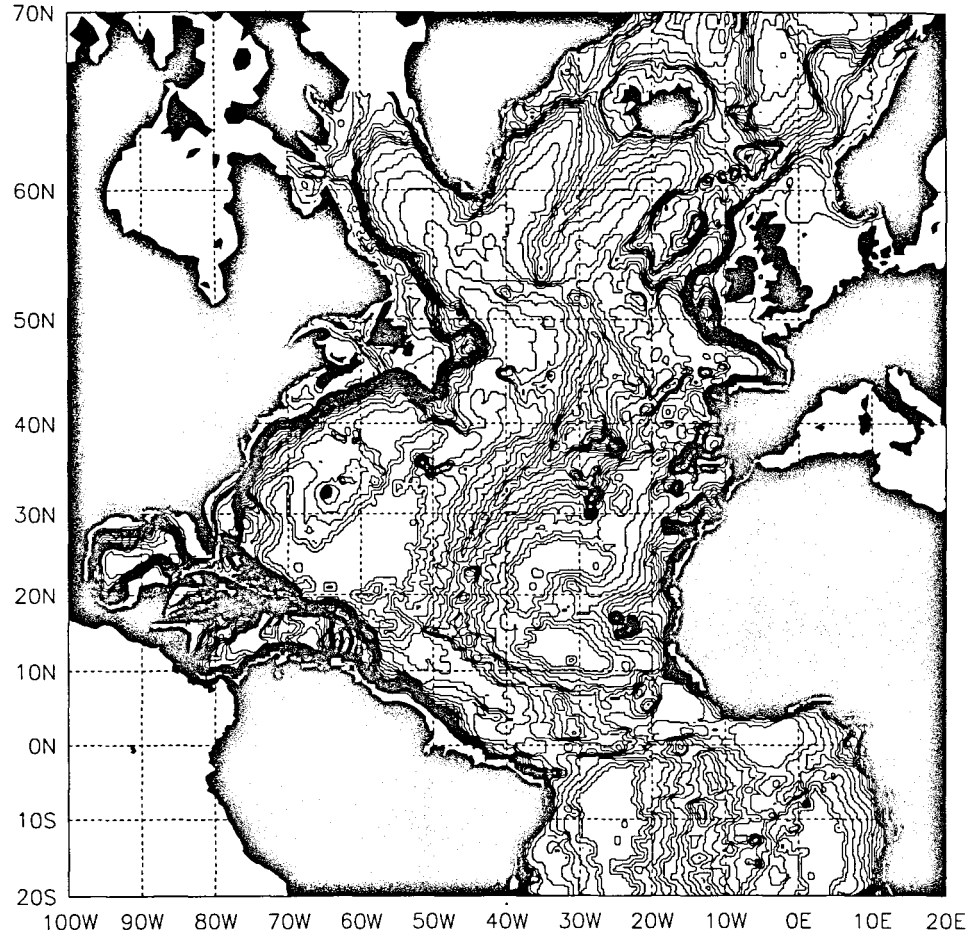


Figure 2.1: LEVEL model bathymetry (m). Contour interval is 100 m above 500 m depth, 250 m underneath.

and a grid-step in latitude, $\Delta\phi$, which varies as the cosine of latitude:

$$\Delta\phi = \Delta deg \cos(\phi) \quad (2.2)$$

Hence the latitude, ϕ_j , of a grid point referenced by the integer index j can be exactly calculated with the following analytical formula:

$$\phi_j = \frac{180}{\pi} \arcsin \left[\tanh \left(\Delta deg \frac{\pi}{180} (j - j_{eq}) \right) \right] \quad (2.3)$$

where j_{eq} is the reference index for the equator.

In addition, the Grenoble and Southampton groups have run coarse resolution versions of SIGMA and ISOPYCNIC, respectively, with a resolution of $4/3^\circ$ at the equator. The coarse resolution version of LEVEL has been used mainly to evaluate different mixed layer parameterisations, and is now used for coupled ecosystem studies with increased vertical resolution by DENG, Princeton University (pers. comm.).

Considerable effort went into achieving a smooth and realistic coastline, particularly around the European margins, and five islands were included, being Iceland, Ireland, Mainland Great Britain, Cuba and Hispaniola. To illustrate the extent of the model domain, the coastline and the rich variability, figure 2.1 depicts as an example the bathymetry of LEVEL.

2.2 Vertical coordinates and bathymetry

The vertical discretisation makes up the main difference of the three models and has a strong impact on the treatment of topography therein. A schematic sketch of the different vertical coordinate systems is given in figure 2.2. Within their respective vertical discretisations, all

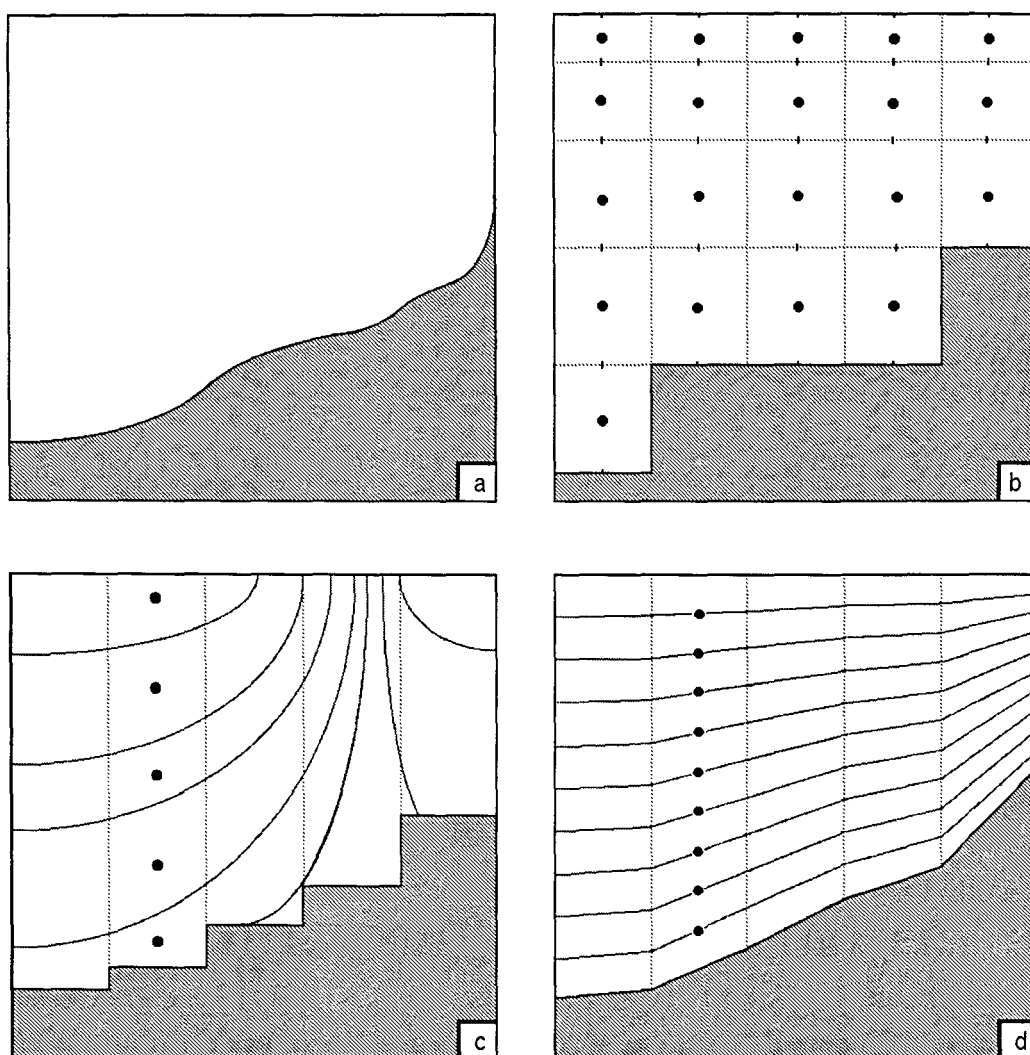


Figure 2.2: Schematic view of the vertical coordinate systems of the three DYNAMO models. (a) original bathymetry section, (b) LEVEL model, (c) ISOPYCNIC model, (d) SIGMA model.

three models use standard staggered grids, with vertical velocities at the upper and lower faces of the cell boxes, and other variables in between. For all models, the maximum depth of the ocean is set to 5500 m, but each model has its own minimum depth, chosen according to its own topographic constraints.

LEVEL uses a step-like representation of the bathymetry, with 30 degrees of freedom, and a grid spacing smoothly increasing from 35 m at the sea surface to 250 m below 1000 m depth. Depths of the vertical levels are shown in Table(2.1). The minimum depth is 72 m, resolved by 2 levels, and the deepest level is centered at 5375 m, yielding a bottom at the base of the grid-cell at 5500 m. Vertical coordinates in LEVEL are identical with the CME model (e.g. DÖSCHER et al., 1994).

Level	Depth (m)	Level	Depth (m)	Level	Depth(m)
1	17.50	11	721.47	21	3125
2	53.43	12	900.87	22	3375
3	91.64	13	1125	23	3625
4	133.19	14	1375	24	3875
5	179.52	15	1625	25	4125
6	232.60	16	1875	26	4375
7	295.03	17	2125	27	4625
8	370.21	18	2375	28	4875
9	462.50	19	2625	29	5125
10	577.37	20	2875	30	5375

Table 2.1: Vertical levels for tracer points in the DYNAMO 1/3° LEVEL-model.

SIGMA uses a vertical discretisation following the bathymetry. The concept of such a vertical coordinate was first introduced in a meteorological model by PHILLIPS (1957), who defined a *sigma* coordinate, σ , as a function of $z/h(x, y)$, where x, y, z are the usual geopotential coordinates and $h(x, y)$ is the local depth of the fluid. The term *sigma* has been kept in oceanographic applications of this coordinate, and may sometimes induce a confusion with potential density σ_θ . In the DYNAMO version of the SIGMA model, σ is a non-linear, analytical function of $z/h(x, y)$ which allows users to increase the resolution near the surface and the bottom, according to the desired application (SONG and HAIDVOGEL, 1994, P. DE MIRANDA, 1996). SIGMA uses 21 levels for vertical velocity (20 levels for tracers), with an increased resolution near the surface as shown in Table(2.2). The coarse vertical discretisation of SIGMA in

deep abyssal plains indicates that it may mis-represent the deepest water masses.

Level		Depth (m)	Depth (m)	Depth(m)	Depth(m)
(surface)	0	0.00	0.00	0.00	0.00
	1	10.00	13.03	20.59	30.05
	2	20.00	27.03	44.62	66.60
	3	30.00	42.66	74.30	113.84
	4	40.00	60.81	112.85	177.89
	5	50.00	82.20	164.78	267.27
	6	60.00	110.27	235.93	393.01
	7	70.00	145.05	332.67	567.19
	8	80.00	188.52	459.82	798.94
	9	90.00	240.63	617.20	1087.91
	10	100.00	299.14	796.98	1419.28
	11	110.00	360.11	985.37	1766.95
	12	120.00	419.74	1169.11	2105.81
	13	130.00	476.29	1342.02	2424.19
	14	140.00	530.52	1506.81	2727.18
	15	150.00	585.02	1672.57	3032.00
	16	160.00	643.32	1851.62	3361.99
	17	170.00	709.36	2057.77	3743.28
	18	180.00	787.45	2306.07	4204.35
	19	190.00	882.42	2613.47	4777.29
(total depth)	20	200.00	1000.00	3000.00	5500.00

Table 2.2: Distribution of the vertical levels for vertical velocity points in the DYNAMO 1/3° SIGMA-model for several values of the total depth. Tracer points are approx. located between vertical velocity points.

ISOPYCNIC uses a vertical coordinate system very different from the other two models. Instead of solving the physical system at selected depths, it uses a set of isopycnal surfaces separating 20 layers of constant densities. The model solves equations of evolution for the usual dynamical and thermodynamical variables, and for the thickness of every layer, under the hypothesis of homogeneity within layers. The set of 20 layers is defined by potential density values referred to the surface, σ_0 , varying from 24.02 to 28.12. These values, shown

in Table(2.3), aim to represent watermasses and thermocline dynamics as close as possible. From the point of view of vertical representation, there arises no need to discretise or smooth bathymetric data.

Layer	σ_0	Layer (cont.)	σ_0 (cont.)
1	24.02	11	27.52
2	24.70	12	27.64
3	25.28	13	27.74
4	25.77	14	27.82
5	26.18	15	27.88
6	26.52	16	27.92
7	26.80	17	28.00
8	27.03	18	28.06
9	27.22	19	28.09
10	27.38	20	28.12

Table 2.3: Potential density, referenced to the surface, of the 20 layers defining the vertical discretisation of the DYNAMO 1/3° ISOPYCNIC-model

Among the differences introduced by the different vertical coordinates, it is worth to mention that SIGMA has the same degrees of freedom in the vertical at any horizontal gridpoint, whereas in LEVEL the number of active vertical levels varies with the horizontal position according to the discretisation of bottom topography. In ISOPYCNIC the thickness of a layer may tend to zero to represent the outcropping of an isopycnal at the surface or its intersection with the bottom topography.

According to their respective coordinate systems, all three models use a slightly different bathymetry. The ETOPO5 database from the National Geophysical Data Center provides the common platform.

For ISOPYCNIC, a simple interpolation onto the model grid by taking the median of all bathymetric data within each grid box, without additional smoothing was performed to prepare the topography. The minimum depth of the ocean is set to 75 m, similar to LEVEL.

Additional effort was necessary to adapt the bathymetry to the discrete vertical grid of LEVEL. To remove the noise on the gridscale, a simple second order SHAPIRO filter was applied to the data interpolated onto the model's horizontal grid. Adoption to the levels was

performed by a nearest point algorithm. To maintain cross sections and sill depths in key regions, handtailoring of the bathymetry was necessary, mainly by using the unfiltered data or by artificially widening straits. The latter is caused by the layout of the ARAKAWA B-grid with the implicit no-slip boundary conditions, where at least two adjacent tracer grid cells are required to allow for an advective transport across a strait.

In SIGMA, the discrete form of the pressure gradient terms may produce systematic errors over steep topography. Thus, a comparatively strong smoothing of the bathymetry is required to control the numerical accuracy of the calculation of the pressure gradient. Various smoothing criteria have been tested in preparation of the bathymetry in SIGMA, expressed as a maximum value of a smoothing coefficient r which combines the vertical and horizontal resolutions with the bottom slope in a way defined by BARNIER et al., (1997):

$$r = \frac{h_i - h_{i-1}}{h_i + h_{i-1}} < \min \left[\left(\frac{\Delta\sigma}{\sigma - 1} \right), 0.2 \right]$$

where h_i is the thickness of the vertical gridcell centered at σ level i . Smoothing in SIGMA has lasting effects on the width of continental shelves and straits, and on sill depths. Thus, the bathymetry has been “re-shaped” empirically in several key areas, such like the Florida Strait and the Bahamas Banks, the Denmark Strait, and the Iceland-Faeroe Ridge system. The minimum depth is set to 200 m to limit the constraint on the time-step introduced by the convergence of the 20 levels onto shallow areas.

As an example, figure 2.3 depicts the three different bathymetries in the Iceland-Faeroes Ridge system, one of the regions of key influence on the structure of the basin-scale circulation.

2.3 Model parameterisation

2.3.1 Lateral mixing and bottom form drag

All models are eddy-resolving, and thus the physical process of mixing by eddies is explicitly resolved. There may be implicit horizontal, isopycnal or along sigma mixing built into the models due to the implicit diffusion of the various discrete numerical schemes. However, the numerical codes have been built to minimize such implicit numerical diffusion of properties, and due to this reason, all models require a representation of subgridscale phenomena.

Both LEVEL and SIGMA use similar biharmonic lateral viscosity and diffusivity acting along geopotential (iso- z) surfaces. In SIGMA this required the rotation of the biharmonic operator

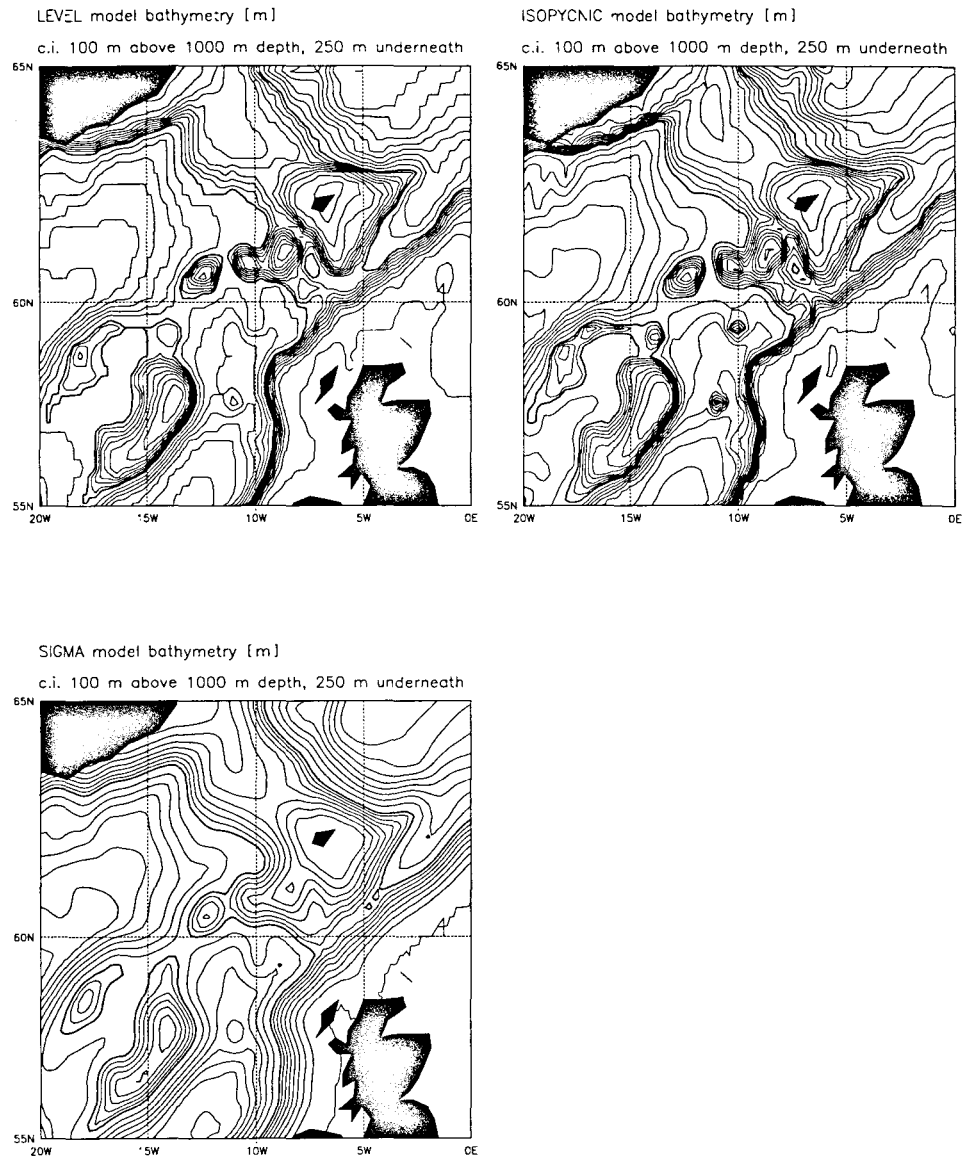


Figure 2.3: Bathymetry of the three DYNAMO models in the Iceland-Faeroe-Shetland Ridge region. Contour interval is 100 m above 1000 m depth, 250 m underneath. (a) LEVEL model, (b) ISOPYCNIC model, (c) SIGMA model.

from sigma to horizontal surfaces (BECKMANN, 1996, pers. comm.). Coefficients are varying with latitude, such that the amount of friction is the same on a given length independent from horizontal position. The spinup was started with diffusivity and viscosity set to $A_{h,m} = -2.5 \cdot 10^{-11} \text{ m}^4/\text{s}$ at the equator.

ISOPYCNIC uses harmonic mixing. There are three mixing coefficients for isopycnic diffusion of momentum, layer thickness and tracers. They are written in the form of diffusion

velocities, and represent the ratio of the diffusion coefficients to the model grid spacing and are constant in the model. For the diffusion of momentum, there is also a shear dependent term, which is effective only in regions of high shear, and is estimated as the product of a viscosity parameter with a deformation velocity calculated from the horizontal shear (BLECK and al., 1992, DYNAMO, 1994).

Parameters are set to

thickness diffusion	=	0.1	(2.0)	cm/s
tracer diffusion	=	0.5	(1.0)	cm/s
diffusion of momentum	=	0.5	(2.0)	cm/s
viscosity parameter	=	0.25	(2.0)	

where values in brackets apply to the coarse resolution model.

It is easily shown that on a length scale of 50 km, the biharmonic eddy viscosity initially chosen for SIGMA and LEVEL corresponds to a harmonic friction velocity on that length scale of 0.2 cm/s, less than the 0.5 cm/s used in ISOPYCNIC. However, it is likely that ISOPYCNIC will be more diffusive of mesoscale features than the other two models, since the harmonic operator is not as scale selective as the biharmonic (HOLLAND, 1978).

Bottom friction in all three models is applied according to a quadratic law with a tidal residual:

$$F = c_d \cdot U \cdot \sqrt{25 \cdot 10^{-4} \text{ m}^2/\text{s}^2 + U \cdot U}$$

where c_d is set to $1.2 \cdot 10^{-3}$, and the velocity at the bottom U is given in m/s.

2.3.2 Diapycnal mixing

All three models share the same vertical (for LEVEL and SIGMA), or diapycnal (for ISOPYCNIC) mixing scheme for tracers, adopted from CUMMINS et al. (1990).

The diffusion coefficient K_v is a function of stratification a_0/N in case of static stability, where N is the local Brunt-Väisälä frequency and a_0 is set to $10^{-7} \text{ m}^2/\text{s}^2$ according to LEDWELL et al. (1993). In case of static instability, the coefficient is set to a value at least 4 orders of magnitude larger than values obtained in stable situations. In ISOPYCNIC, unstable stratification never occurs away from the influence of the surface forcing. A convective adjustment, which mixes temperature, salinity and momentum, takes place only between the mixed layer and the isopycnal layers below it when the mixed layer density exceeds the prescribed isopycnal layer densities.

In LEVEL and SIGMA, the vertical mixing of momentum is carried-out by an implicit harmonic diffusion, with a constant coefficient $A_{v,m} = 10^{-3} \text{m}^2/\text{s}$. There is no diapycnal mixing of momentum between layers in ISOPYCNIC.

2.3.3 Mixed layer concepts

In the near-surface layers of the ocean, vertical mixing is enhanced due to turbulent motion produced by wind, waves, current shears or surface cooling and evaporation. The turbulent kinetic energy generated by these processes is converted to potential energy of the stratification by mixing the light surface water with denser water from below.

Although this turbulence is not resolved by the DYNAMO models, the production of turbulent kinetic energy and its effect on the near-surface mixing can be parameterised by use of an explicit mixed layer sub-model.

The conceptually most simple model of the oceanic mixed layer is the Kraus–Turner bulk model (KRAUS and TURNER, 1967; NIILER and KRAUS, 1977), in which the ocean's properties are assumed to be completely homogenised within the mixed layer. Instead of a smooth transition to the values below the zone of intensified mixing, the model assumes a sudden jump in velocities and tracer concentrations at the base of the mixed layer. The generation of turbulent kinetic energy is parameterised by relating it to the forcing functions, such as the strength of the wind or the buoyancy flux, and the time rate of change of mixed layer depth is then determined from the turbulent kinetic energy equation. Although this model may be oversimplified in some respects, it simulates the observed mixed layer evolution reasonably well (cf. GASPAR, 1988; VUILLEMIN, 1995) while at the same time allowing a basic understanding of the processes involved.

The mixed layer concept retained in DYNAMO is based on the Kraus–Turner model. Basic mechanisms, numerical considerations, and choice of parameters have been extensively discussed in the second DYNAMO scientific report (1995), and in an informal report from DENG (pers. comm., 1995).

Mixing of tracers

For the intercomparison experiments, only ISOPYCNAL uses a full Kraus–Turner type of mixed layer model, which description can be found in BLECK et al. (1989) and BLECK et al. (1992). The choice of parameters relative to the implementation of the Kraus–Turner scheme in ISOPYCNIC is described in a DYNAMO scientific report (1995). LEVEL and SIGMA both use a mixed

layer of constant depth, in the sense that the forcing is applied to the the first grid cell in LEVEL and to a 50 m depth body force in SIGMA, associated with a convective ajustement of the tracer fields based upon a test of static stability of the water column. Therefore, for these two models, the mixed-layer depth is not a direct ouput of the model calculation; it has to be diagnosed from the properties of the buoyancy fields with a criterion to be defined. The criterion retained in DYNAMO is an increase in potential density of 0.01 between the surface and the base of the mixed layer.

Sensitivity experiments carried out with LEVEL use a full Kraus-Turner mixed layer model, a detailed description of which can be found in a report by DENG (pers. comm., 1995).

The type of mixed layer parameterisation used in every DYNAMO experiments is presented in Table(2.4) with other characteristics. The differences pointed-out above may later be of interest in the interpretation of the model results. Whichever is the mixed layer scheme used (constant or Kraus-Turner), tracers are completely homogenised within the mixed layer in all three models.

Mixing of momentum

One major difference between ISOPYCNIC and the other two models rests in the mixing of momentum. Due to the formulation of the model equations on isopycnal layers, the mixed layer (i.e. the only place where diapycnal fluxes are permitted) is always confined to the top layer of the model. As this layer can only have one value for T , S and \bar{v} at any geographical position, the velocities in the mixed layer are uniform. For deep mixing, this immediately implies small horizontal velocities because the momentum is spread over the whole layer.

In LEVEL and SIGMA, no homogenisation of momentum takes place in the mixed layer, and strong current shears may persist. This has to be kept in mind when comparing near-surface currents in regions of deep mixing.

Forcing data

In the Krauss–Turner formulation of the mixed layer, it is important to account for the highly nonlinear nature of the input of turbulent kinetic energy (TKE). Therefore, the monthly means of the third power of the friction velocity u_* , required in the TKE–equation as given by NIILER and KRAUS (1977), have been determined from the 6-hour ECMWF analyses by the Grenoble group (cf. DYNAMO scientific report, 1994). This was done in a manner consistent with the wind and temperature forcing used for DYNAMO. The surface buoyancy flux,

Table 2.4: Model configuration and parameters of the CME and DYNAMO 1/3°-models

	Boundaries		Resolution			Lateral friction/diffusion	Mixed layer	Wind forcing
	North	South	Zonal	Merid.	Vertical			
CME	65°N Relaxation or open	15°S Relaxation or open	2/5°	1/3° B-grid	30 levels	biharmonic	Kraus-Turner	various climatologies
ISOPYCNIC	70°N Relaxation	19.6°S Relaxation	1/3°	$1/3^\circ \cdot \cos \phi$ C-grid	20 layers	harmonic	Kraus-Turner	Monthly ECMWF 1986–1988
SIGMA	70°N Relaxation	17.5°S Relaxation	1/3°	$1/3^\circ \cdot \cos \phi$ C-grid	20 levels	biharmonic	constant	Monthly ECMWF 1986–1988
LEVEL – “No Wind” – “Daily Wind”	70°N Relaxation	20°S open	1/3°	$1/3^\circ \cdot \cos \phi$ B-grid	30 levels	biharmonic	a) constant b) Kraus-Turner	Monthly ECMWF 1986–1988 no wind stress Daily ECMWF 1986–1993

required in the formulation used in ISOPYCNIC (BLECK et al., 1989) is derived from the surface relaxation of T and S towards the prescribed surface forcing data.

2.3.4 Initialisation and surface forcing

All models are driven by a forcing function derived from global 6 hourly analyses performed at ECMWF. Analysed fields have been studied by the Grenoble group, and a detailed description of the forcing data is given in Appendices A and B of the DYNAMO (1994) scientific report.

The seasonal atmospheric forcing is a monthly-mean climatology obtained from three years of ECMWF analyses (1986 to 1988). The formulation of the surface heat flux is described in BARNIER et al. (1995) where patterns of the heat flux are presented. A linearization of bulk formulas is used to define a model-dependent air-sea net heat flux which can be applied to the models equivalently as a surface flux boundary condition, as a source term in the equation of temperature in the case of a body force formulation, or a relaxation to an equivalent sea surface temperature. The wind stress is also derived from the same ECMWF analyses (SIEFRIDT, 1994), using the formulation proposed by KONDO (1975). No precipitation field is available from ECMWF for this period. Thus, the fresh water forcing is a relaxation of the model surface salinity to the climatological values provided by Levitus (1982) with a time scale identical to that given by the formulation of the heat flux (BARNIER et al., 1995).

For the response experiments and the assimilation of altimetric data, daily windstresses are provided by the Grenoble group for the period 1986 to 1993, also from ECMWF analyses.

The ECMWF wind stresses, atmospheric equilibrium temperature, friction velocity and Levitus (LEVITUS, 1982) sea surface salinity were all converted to pseudo fields as proposed by KILLWORTH (1996), in order to achieve the correct monthly means. The relaxation timescales for the surface heat and freshwater fluxes were left unchanged. The surface fluxes are applied to the surface layer or mixed layer respectively as described in previous reports.

All models are relaxed to climatological conditions near the northern boundary and near Gibraltar. At the southern boundary, the conditions in the 3 models were however different. ISOPYCNIC and SIGMA have an additional restoring zone adjacent to the southern boundary, whereas for LEVEL a presumably more realistic open boundary condition has been developed which basically follows the algorithm developed by (REDLER and BÖNING, 1997) for the CME-model. Thus for LEVEL the relaxation occurs only on inflow points, and in addition the normal component of the barotropic velocity is prescribed according to the Sverdrup relation. A preliminary investigation of surface drifter data (cf. SCHÄFER AND KRAUSS 1995) has shown

that at 20°N near-surface velocities are nearly zonal, hence the different boundary conditions should not result in large model differences.

For the area near the northern boundary, data supplied by the Kiel group was merged into the Levitus data to provide an improved hydrographic database for the relaxation. For the other areas, the usual Levitus dataset was used. The restoring data have been provided by the Kiel group on a regular $1/3^\circ$ grid using the LEVEL coordinates in the vertical.

For ISOPYCNIC, the Kiel datasets (T and S) were first interpolated vertically on to the standard Levitus levels, then interpolated horizontally onto a subset of the model grid (north of 60°N, and east of 40°W) by the quasi-hermite method described in the 1994 Scientific Report. On the model grid, Levitus T and S in the north-east portion of the domain were then replaced by the Kiel data. Specifically, a 175 (east-west) \times 72 (north-south) grid mesh in fine resolution ISOPYCNIC in the north-east corner has Kiel data (a 45 by 18 grid mesh for the coarse model). The merged datasets (T and S for 12 months) were then convectively adjusted and interpolated onto density layers as described in the 1994 Scientific Report, and then used for the initialisation and lateral relaxation of the model. Below the base of the mixed layer, relaxation applies to layer salinity and interfaces depths. In the mixed layer, potential temperature and salinity are relaxed to the potential temperature and salinity of Levitus (or the Kiel data near the northern boundary) averaged over the depth of the mixed layer. Full depth relaxation applies to the Northern and Southern boundaries, but only the top 14 layers have lateral relaxation near Gibraltar.

For all models, north of 67°N from 40°W to 10°W (over 24 grid points in the north-south direction in the fine resolution configurations), the relaxation time scale linearly increases with distance from 3 days at the northern boundary to 100 days at approximately 67°N. East of 10°W, the relaxation zone is north of the straight line connecting (10°W, 67°N) to 60°N at the model eastern boundary. The relaxation time scale again increases from 3 days at the northern boundary to 100 days at the southern edge of the zone. The Strait of Gibraltar in the model is defined at approximately 36°N and 6°W. The area of relaxation is a 16 \times 16 grid mesh with the Strait of Gibraltar in the centre of the eastern side of the mesh. The time scale is 14 days at the Strait, and increases linearly with distance to 100 days at a distance of 300 km and thus reflects the character of a point source for the Mediterranean Water.

All three models include a “zero order ice model” with a slightly different formulation. In LEVEL, if the predicted surface or mixed layer temperature is below the temperature for sea ice (-1.8°C), then ice cover is assumed, and the surface heat flux, salinity flux and friction velocity are all set to zero. In ISOPYCNIC and SIGMA, ice cover is assumed when the forcing

surface temperature is below the freezing temperature for sea ice (-1.8°C). Then, surface heat flux is set to zero, and the model surface temperature is relaxed to the temperature of sea ice with a 3 day time constant.

The spinup of all three models starts from a state of rest using the Levitus climatology for the month of September, when the ocean is well stratified.

The essential physical parameters of the three competitive models are compiled in table (2.4)

Chapter 3

Model Experiments and Analysis

3.1 Prognostic model experiments

The backbone of the project is provided by integrations of the three 1/3-degree model versions LEVEL, ISOPYCNIC and SIGMA over an identical, 20-year period, starting from the initial conditions given by the Levitus-climatology. The last 5 years of these runs serve as the main analysis period for the model intercomparisons and evaluations. The choice of a 15-year spin-up period is motivated by the experience from previous model studies of the North Atlantic which showed that the velocity field reaches a dynamical, quasi-equilibrium state 10–15 years after the application or change of the external forcing. The time scale of this dynamical response is basically set by the passage of the lowest mode baroclinic Rossby wave across the basin, both with regard to the wind forcing at the surface (ANDERSON et al., 1979), and to the thermohaline forcing at the surface and the northern and southern boundaries which is responsible for the spin-up of the meridional overturning cell (DÖSCHER et al., 1994; GERDES and KÖBERLE, 1995). Figure 3.1 shows time series of the overturning intensity at 43°N, indicating only little drift in the model transports after 10 years of integration.

Any model intercomparison has to take into account a possible, but often unknown dependency of the different models on a considerable number of model choices: details of the model topography, boundary conditions or mixing parameterisations can all have significant impact on the behaviour of the basin-scale circulation. Ideally, a comparison of different numerical models with regard to their ability to reproduce observed features of the ocean should not be based on single realisations of these models, with a given set of parameters, but between sets of experiments that give an idea of their parameter dependencies. Since, of course, this is far beyond present computational resources, we need to take into account,

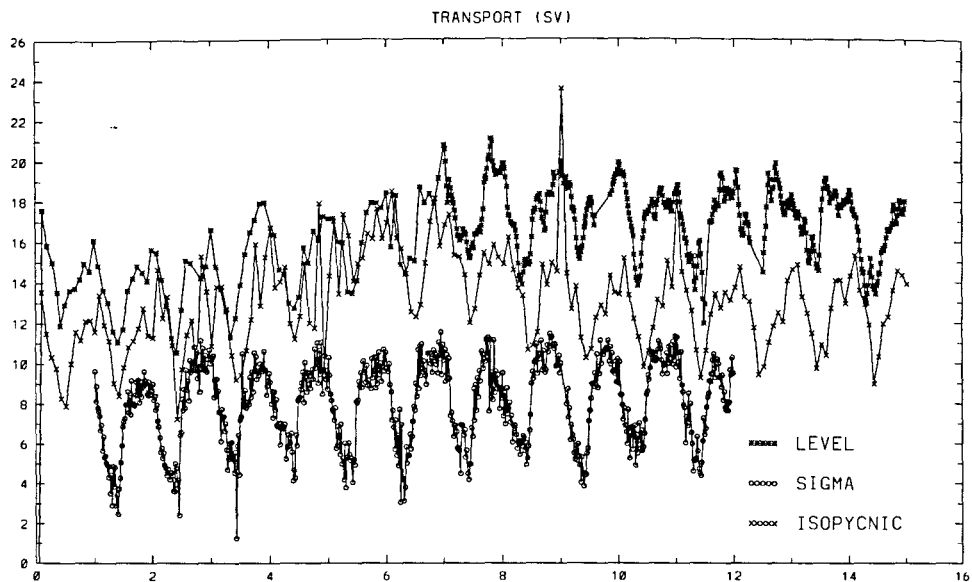


Figure 3.1: Temporal evolution of maximum overturning (Sv) at 43°N during the spinup of the three DYNAMO models.

as far as possible, the information available from sensitivity studies with similar model configurations. These have, in particular, been carried out under the CME framework where a suite of model versions differing in horizontal resolution, friction, wind forcing and thermohaline boundary conditions were examined, and may be utilised to assess the solution of the DYNAMO LEVEL model. Some experience also exists with previous North Atlantic applications of the Miami isopycnic model for the North Atlantic, including comparisons between non-eddy resolving versions of that model with similar versions of the GFDL model (MARSH et al., 1996; ROBERTS et al., 1996).

In DYNAMO we have tried to add to the understanding of model dependencies by including a sensitivity experiment focussing on the effect of a Kraus-Turner mixed layer scheme for wind-induced mixing in LEVEL. Figure 3.2 gives a graphical overview of the different DYNAMO experiments, in the context of previous sensitivity and intercomparison studies with North Atlantic models.

3.2 Model analysis, intercomparison and validation strategy

The analysis of a high-resolution, basin-scale ocean model is a challenging task due to the diversity of physical processes on a wide range of space and time scales comprised in the solution. The integration originally produces a succession of three-dimensional fields of a set

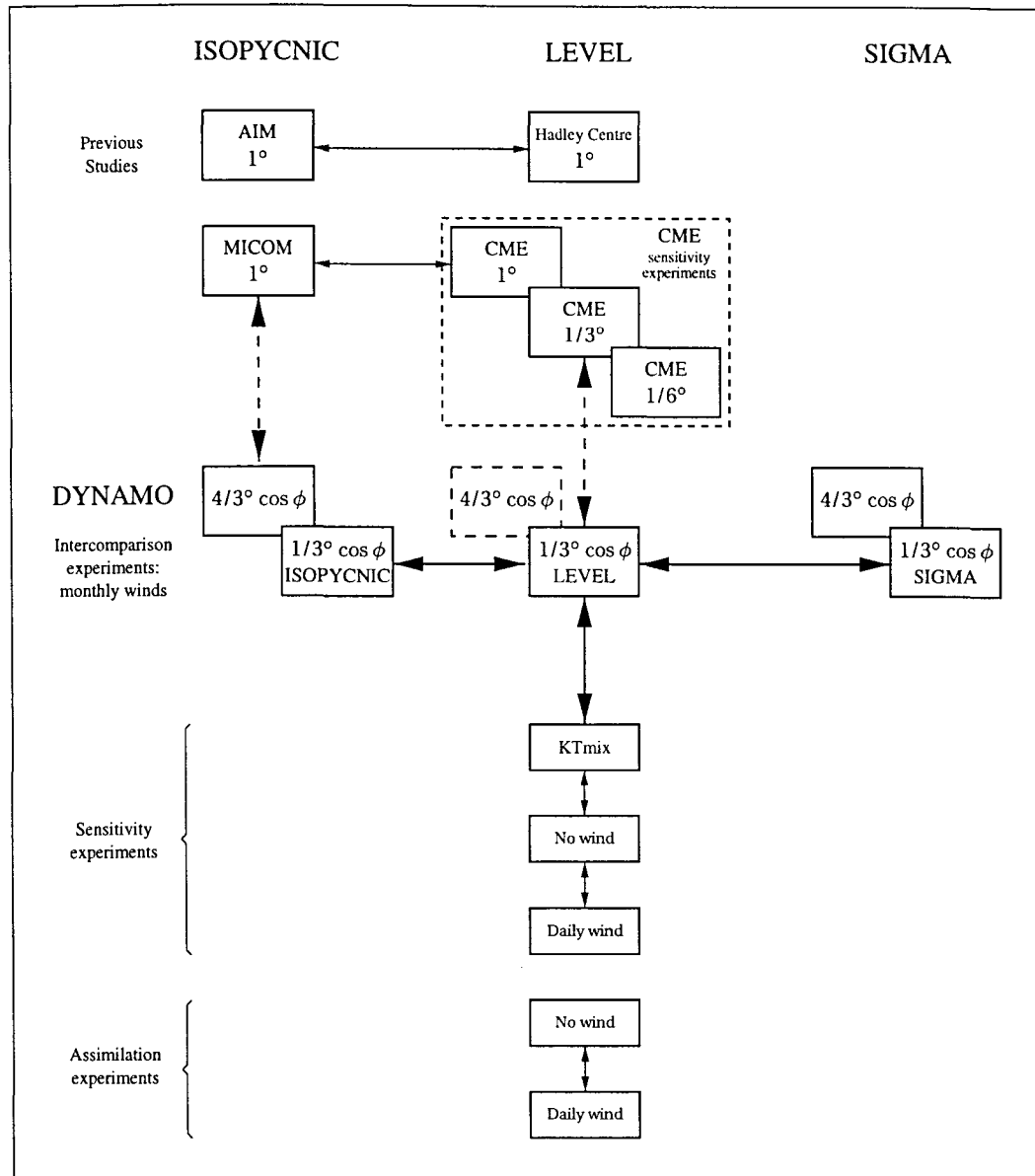


Figure 3.2: DYNAMO experiments, in relation to previous model studies and intercomparisons for similar North Atlantic configurations.

of variables: the prognostic variables of the respective model, plus a number of fields diagnosed during the integration such as heat and freshwater fluxes or sea surface height. Even if it were possible to store all these fields on a modern mass-storage system – with a sampling rate of a few days necessary for assessing the eddy variability –, the retrieval of information for different analysis purposes would be a rather daunting task and, for many applications, prohibitively time-consuming. Any practical strategy for the storage of model output therefore has to be a compromise between comprehensiveness and accessibility, tailored as much as possible towards the key scientific issues, but without excluding additional analyses at later

stages. We will here briefly describe the main strategy for the storage of the output from the prognostic DYNAMO runs, and give an overview of the specific data sets stored in a common format by all groups.

The focus of the DYNAMO analysis is on an evaluation of the three numerical models with respect to their mean circulation, seasonal variation, and eddy statistics for the 5-year period following the 15-year spin-up. For the model intercomparison, three dimensional quarterly mean fields of a host of model variables and eddy correlation terms have been compiled, generally interpolated on to a common A-grid consisting of the mass-point coordinates in the horizontal and 61 non-equidistant levels in the vertical. In order to explicitly account for the temporal variability of the model flows, all groups had agreed upon a number of temporal slices for selected individual points, cross-sections, and integral quantities. For the spin-up phase the main requirement was to document (i) the drift of the models' hydrographic properties from the initial state, (ii) the spin-up and dynamical adjustment of the circulation from the state of rest. For the intercomparison phase the main idea was to extract information for a number of sections which would allow direct comparison with observations, e.g., transport sections for boundary currents and the overflows.

For additional diagnostics, but more time-consuming to assess, all groups have stored full snapshots of the basin-scale fields at monthly intervals during the spin-up phase, and at three day intervals for most of the intercomparison and response experiments.

The total inventory of DYNAMO-model output amounts to roughly a terabyte of data, and will serve as a basis for ongoing analysis after the end of the project.

In the following, we give an overview of the specific data sets compiled for the spin-up and intercomparison phases.

3.2.1 Snapshots and time series for the spin-up phase

During the spinup, timeseries are stored with a sampling rate of 6 days to monitor the models' drifts from the initial state and the spinup of the circulation from the state of rest. With respect to water mass properties and thermohaline circulation, timeseries of horizontally averaged temperature and salinity on selected horizontal surfaces are stored, as well as the heat and salinity content in the mixed layer, the volume of selected water masses (Sub-Tropical Mode Water, Sub-Polar Mode Water, Mediterranean Water, upper and lower North Atlantic Deep Water, and Antarctic Bottom Water), profiles of the meridional overturning streamfunction at selected latitudes, meridional (zonally and vertically integrated) heat and salt transport at selected latitudes, and profiles of temperature and salinity at distinct locations (i.e.,

in North East Atlantic, Labrador Sea, Central Subtropical Gyre, at the Equator). Additionally, instantaneous maps of the depth of the mixed layer, its density and temperature on March 15 and September 15 are stored.

With respect to the overflow problem, binned transports for selected density ranges normal to a couple of sections have been stored (e.g. across Iceland-Scotland Ridge, Denmark Strait, East Greenland Current, a north-south section along 44°W from the Greenland coast to 47°N , zonal sections along 25 and 43°N).

For further diagnostics, monthly snapshots have been stored during the spinup phase.

3.2.2 Data sets for intercomparison phase and sensitivity experiments

The main data sets for the evaluation of the intercomparison and response experiments are the five-year mean, three dimensional fields for each season and the annual mean, compiled with a sampling interval of 3 days.

With respect to water mass properties and thermohaline circulation, this dataset comprises a basin-wide water mass census in classes of 0.1 K and 0.05 psu , meridional overturning streamfunction (versus depth and potential density), maps of correction heat and fresh water fluxes, meridional sections of density and large scale potential vorticity along 30°W and a North Atlantic Current section along 48°N .

Volume Transports are computed from all prognostic variables stored along additional cross sections (e.g. DICKSON, 1983)

Eddy statistics are based on maps of sea surface height and its variability, and climatological quarterly means and correlations of almost all prognostic and diagnostic variables (e.g. $u, v, \theta, S, \rho, q, uu, vv, \theta\theta, SS, \rho\rho, qq, u\theta, v\theta, w\theta, uS, vS, wS, u\rho, v\rho, w\rho, uq, vq, wq \dots$).

With respect to certain regional aspects of the circulation, additional subsets of data have been stored during the experiments, and even though not explicitly agreed upon, all three groups managed to store full three day snapshots for most of the intercomparison and response experiments.

Chapter 4

Basin-Scale Overview

4.1 Thermohaline circulation

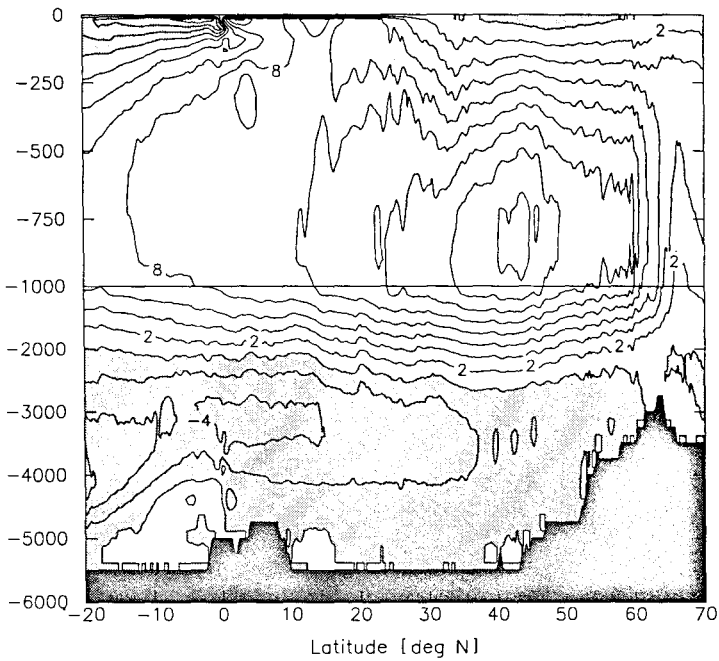
The North Atlantic is the region where most of the world ocean's deep water is formed. The ability of the three models to reproduce essential features of the thermohaline circulation is therefore an important aspect of their performance.

All three models have been integrated for a total of twenty years, a short time compared to the thermohaline response time of the deep ocean. Hence, while the circulation field is well-adjusted to the density distribution, the water mass distributions are far from equilibrium and reflect the initial state to a considerable degree. Nevertheless, a substantial adjustment already occurs on decadal timescales which allows a meaningful comparison of the thermohaline circulation in the three models.

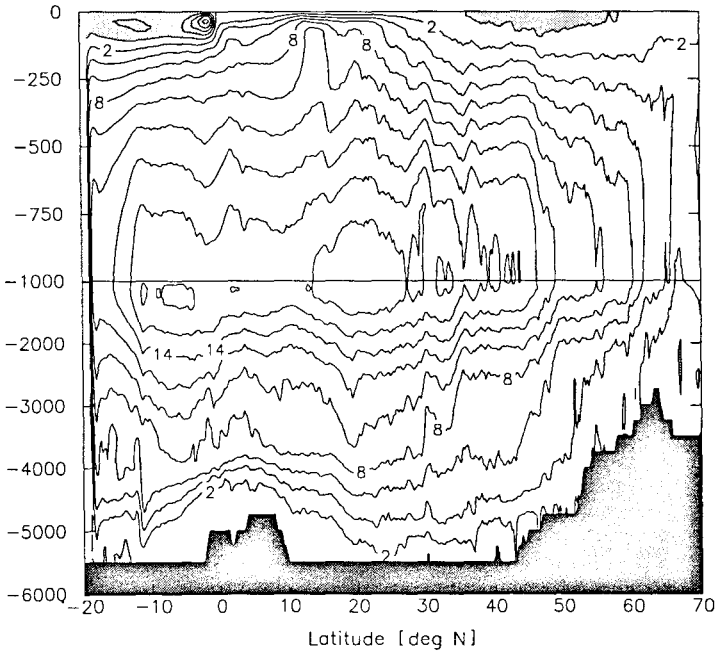
4.1.1 Meridional overturning

The zonally integrated transport is the variable most frequently used to characterise the strength of the thermohaline circulation although it can neither be observed in a direct way nor is its interpretation unambiguous.

The stream functions of the zonally integrated transport (Fig. 4.1) show that the overall strength of the thermohaline cell is fairly similar in all three models, with a maximum transport of 16–20 Sv. The spatial structure of the thermohaline cells is however distinctly different, implying different pathways for the North Atlantic Deep Water (NADW). The overflow of NADW across the northern ridges in LEVEL (Fig. 4.1(a)) amounts to 4 Sv, and an additional 8 Sv transport occurs north of 60°N through entrainment and/or sinking in the subpolar gyre. The NADW transport amounts to 12 Sv at the latitude of Cape Farewell, and further south the



(a) LEVEL



(b) ISOPYCNIC

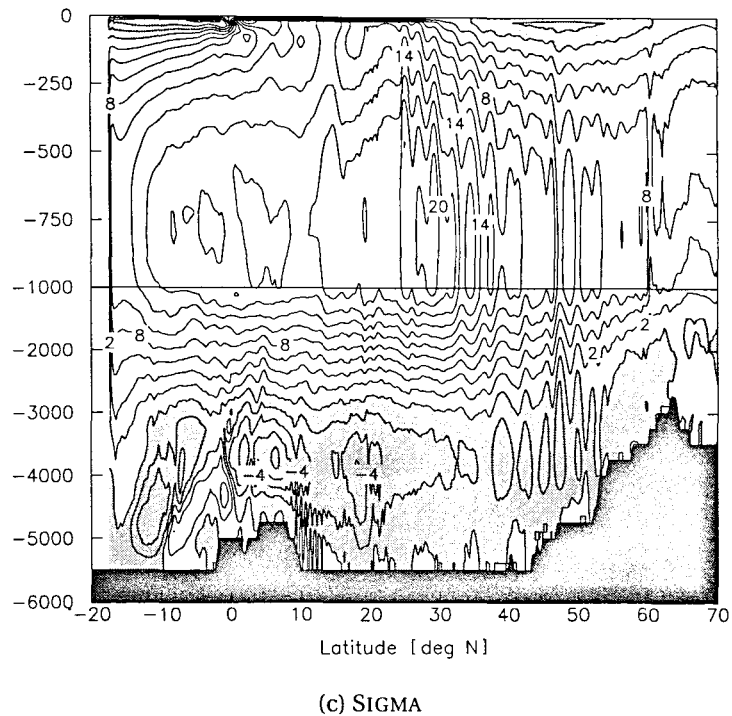
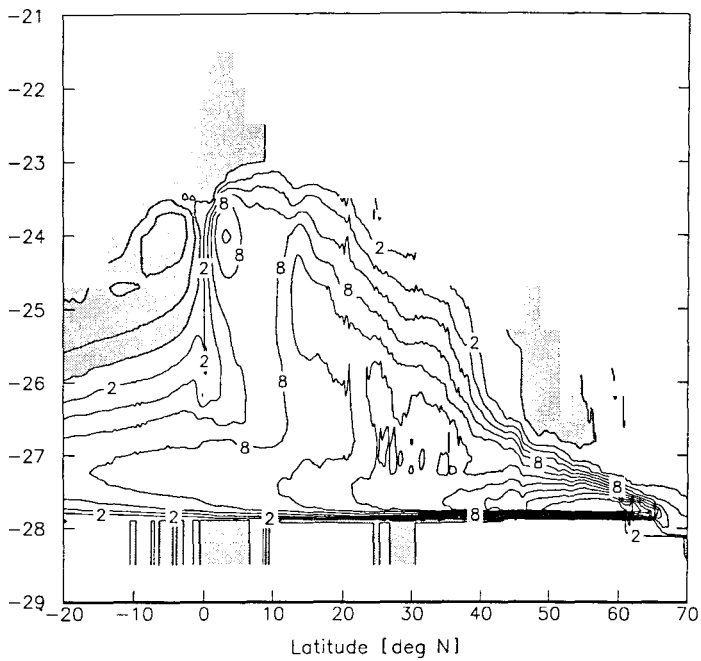


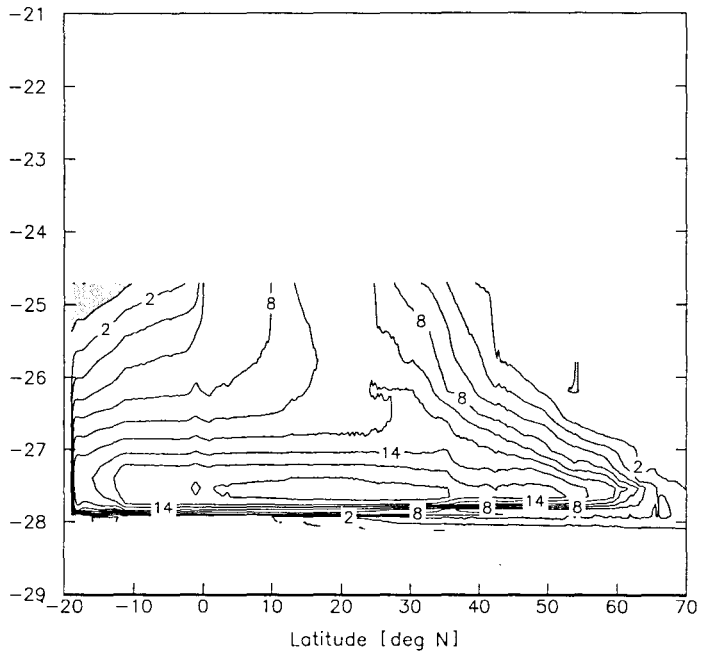
Figure 4.1: Meridional overturning stream function for the three models, plotted vs. depth.

thermohaline cell increases to a maximum of 16 Sv at 40°N and 900 m depth. The overflow transport in ISOPYCNIC (Fig. 4.1(b)) is also 4 Sv, but the total transport at 60°N is only 8 Sv. Further sinking occurs fairly evenly between 60°N and 40°N which originates from the Labrador and Irminger Seas, leading to a maximum overturning of 18 Sv at 20°N and 1,000 m. SIGMA (Fig. 4.1(c)) has the strongest overflow (6 Sv) which is however partly recirculated northwards. The sinking between 65°N and 55°N is weak, and at 60°N the total transport is still at 6 Sv. Additional sinking is concentrated in several latitude bands down to 30°N, and the cell maximum is 20 Sv at 28°N and 800 m. It is noteworthy that, for the first time, basin-scale models have succeeded in simulating an overflow with approximately the correct magnitude without substantial modifications of the bottom topography.

Both LEVEL and SIGMA have deep reverse cells which are somewhat different in strength and pattern, whereas ISOPYCNIC has no reverse cell at all. LEVEL shows a northward flow below 3,500 m reaching 4 Sv which may be interpreted as transport of Antarctic Bottom Water (AABW). The reverse cell in SIGMA occurs at greater depth and is significantly stronger but also shows a number of closed recirculations, making the interpretation more difficult. Overall the differences between LEVEL and ISOPYCNIC at high-resolution are qualitatively similar,



(a) LEVEL



(b) ISOPYCNIC

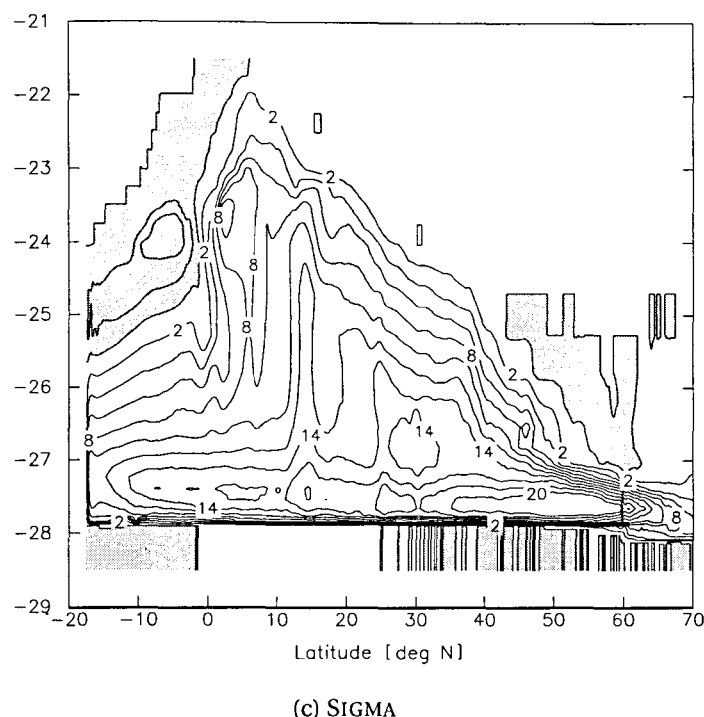


Figure 4.2: Meridional overturning stream function for the three models, plotted vs. density σ_θ .

but substantially less in magnitude, than the corresponding differences in the intercomparison of 1-degree models by CHASSIGNET et al. (1995). In the tropical South Atlantic, transports of 3–5 Sv have been estimated by several authors for the AABW (MCCARTNEY and CURRY, 1993; SPEER and ZENK, 1993 and SPEER et al., 1996). MCCARTNEY and CURRY (1993) estimated a cross-equatorial flow of AABW of the order of 4.3 Sv. This value is in very good agreement with those from the LEVEL and SIGMA models (Figs. 4.1(a) and 4.1(c)).

Another view of the thermohaline transport is given by the zonally integrated transport displayed vs. density, chosen here as σ_θ to facilitate intercomparison of all three models (Fig. 4.2). The water mass transformation between different density classes is generally rather localised in LEVEL and SIGMA, and much more uniform in ISOPYCNIC. The deep transport at the highest densities is very nearly along isopycnals in all three models. The densification in ISOPYCNIC between 60°N and 30°N indicates gradual mixing of Labrador Sea-Irminger Sea water with the deep overflow water, resulting from the explicit diapycnal mixing in the model. In the subpolar region the transport is strongest in LEVEL and SIGMA, but the thermohaline transport in LEVEL is much weaker at low latitudes than that of both other models because of strong diapycnal upwelling between 50°N and 30°N. The reason for this strong upwelling

is unclear. In coarse-resolution models it can be interpreted as spurious upwelling inside the western boundary current, and can be cured by applying the GENT-MCWILLIAMS eddy-mixing scheme (BÖNING et al., 1995). However, LEVEL has an even better resolution than the high-resolution CME model where this upwelling was largely absent, so it is not obvious that this explanation is correct.

The actual overturning transport cannot be observed directly, and indirect estimates differ substantially depending on which method and which observations are used for its determination. Based on the budget of various tracers, a maximum overturning of 20 Sv has been inferred (BROECKER, 1991). HALL and BRYDEN (1982) determined from a geostrophic calculation and measurements of the Florida Current that 18 Sv of water warmer than 7°C are transported northward at 24°N. Based on water mass analysis and geostrophic calculations from hydrographic sections, SCHMITZ and MCCARTNEY (1993) have given a more detailed description of the circulation system, including various recirculation regimes, and obtained the somewhat lower value of 13 Sv crossing the Equator, with little evidence of upwelling between Equator and 50°N. From inverse calculations, RINTOUL and WUNSCH (1991) determined the transport of NADW at 24°N and 36°N to be ≈ 20 Sv, similar to the estimate of 17 Sv obtained by MACDONALD (1995),

4.1.2 Thermohaline transport at 25° N

The different vertical structure of the thermohaline transport at 24°N in the three models is apparent in Fig. 4.3. The northward flow of main thermocline water is above 1,000 m in ISOPYCNIC, and more concentrated above 500 m in LEVEL and SIGMA. The spreading of NADW occurs between 1,000 and 3,500 m in both LEVEL and SIGMA, with a maximum near 2,000 m, followed by northward transport of AABW below 3,500 m which is more pronounced in SIGMA. The structure of the southward flow in ISOPYCNIC is different, with two maxima at 2,000 and at 4,000 m, respectively, corresponding to the upper and lower branches of NADW, and no northward transport at all in the deep ocean. Transport estimates based on inverse calculations from hydrographic section data at 24°N (ROEMMICH, 1985) show that the 1981-section has a clear maximum at 4,000 m associated with lower NADW, whereas in the 1957-section the lower NADW transport is weaker (but still visible). A more recent analysis (MACDONALD, 1995) also suggests that the zonally integrated transport between 1,500 m and 5,000 m is southwards, with a secondary maximum below 4,000 m. A similar pattern prevails at 36°N (Fig. 4.4).

Again both SIGMA and LEVEL fail to reproduce the lower branch of the NADW which is

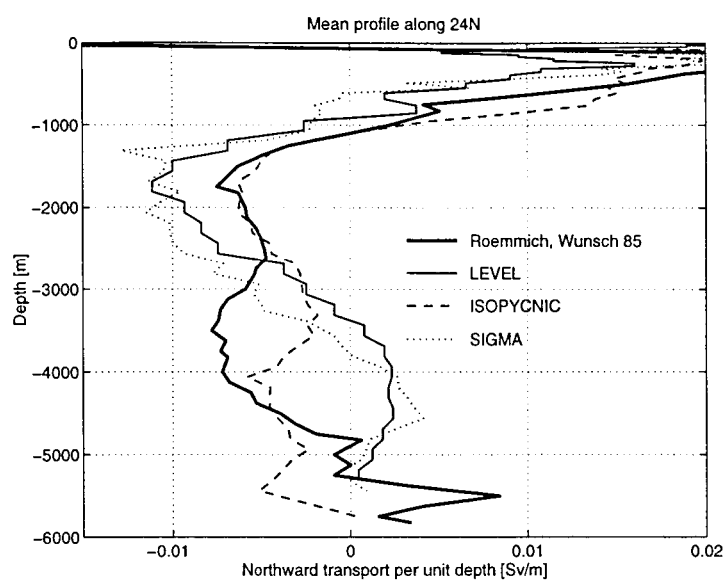


Figure 4.3: Overturning transport per unit depth at 24°N vs. depth for all models.

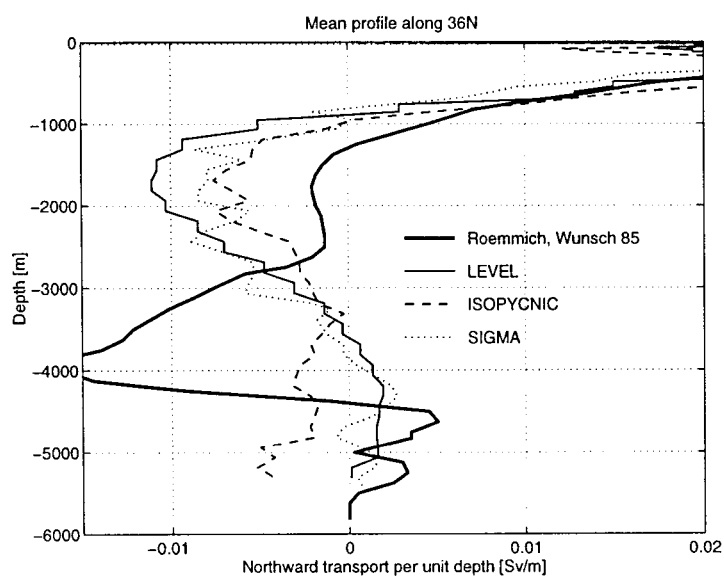


Figure 4.4: Overturning transport per unit depth at 36°N vs. depth for all models.

present in the analyses of ROEMMICH and WUNSCH (1985), MACDONALD (1995) and RINTOUL and WUNSCH (1991). The only model that has a strong southward transport between 3,500 and 5,000 m is ISOPYCNIC which however fails to reproduce the northward AABW transport.

The absence of lower NADW-transport in LEVEL and SIGMA is probably not an inherent defect of these models, it rather reflects deficiencies in deep densities in the northern Irminger and Island Basins, due to strong and localised mixing of the dense overflow water which may be associated with the overflow, and to some degree also with the southern boundary condition. This is suggested by the results of DÖSCHER et al. (1994) who obtained with a level model a pronounced transport maximum below 4,000 m depth when assuring the presence of sufficiently dense deep water through the formulation of the restoring boundary condition which was used to resemble the deep water formation through overflows that were not explicitly included. The density distribution will be discussed in more detail in chapter 5 below.

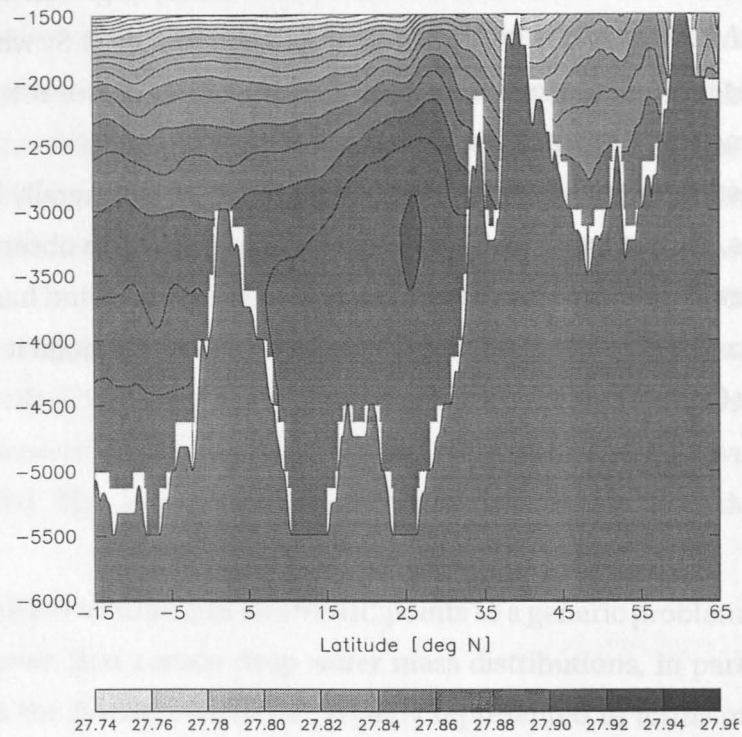
The absence of AABW transport in ISOPYCNIC points to a generic problem in the isopycnal model. It is well known that certain deep water mass distributions, in particular including that associated with the AABW, cannot be properly represented in terms of σ_θ which is the only density variable in the isopycnal model. By the same token, the velocity shear in ISOPYCNIC can differ substantially from its correct value which is connected to the in-situ density rather than to σ_θ . This is exemplified by several isolated maxima which occur both in the climatology and in the other two models in Figs. 4.5 and 4.6 below whereas the distribution in ISOPYCNIC is necessarily monotonous.

This problem cannot be resolved through an increase in vertical resolution, it could however probably be reduced by choosing a density variable that is more representative for the deep circulation although such a choice might lead to deficiencies in the upper part of the water column.

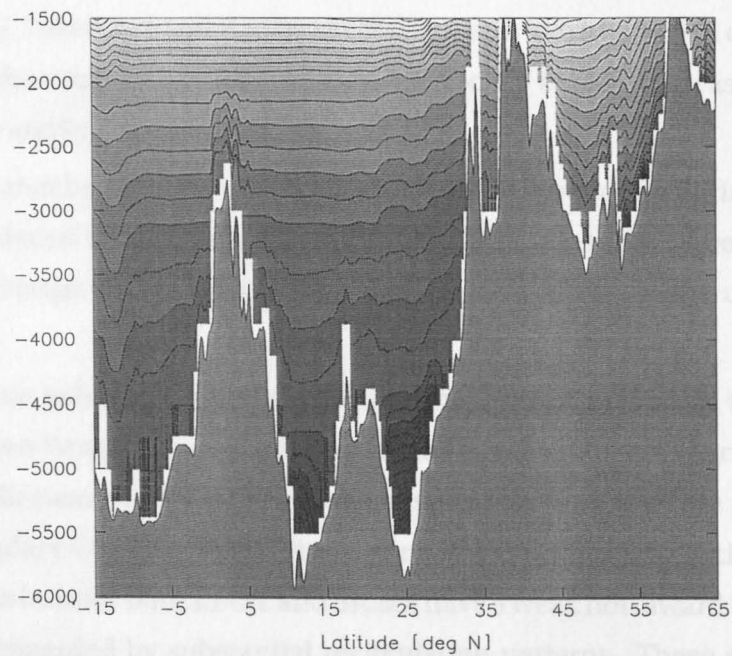
Fig. 4.7 gives some information on how the thermohaline transport in the three models is partitioned between Western Boundary Current and interior. For comparison results from two of the older CME runs as well as from two sensitivity experiments are also shown. The Deep Western Boundary Current (DWBC) transports in LEVEL and SIGMA closely agree at 17 resp. 16 Sv, and in the interior both LEVEL and SIGMA have a weak northward transport (4 resp. 2 Sv) which is accompanied by substantial recirculation patterns. These values agree well with the estimates by SCHMITZ and MCCARTNEY (1993) who find 17 Sv for the DWBC transport between 1.8 and 4°C. The DWBC transport in ISOPYCNIC is somewhat lower at 11 Sv, with an interior transport that is again southwards (9 Sv), accompanied by 32 Sv of recirculation

which occurs mainly in the eastern basin below 4,000 m. While the partition between WBC and interior depends on the exact location of the box boundary (e.g. with a partition at 71°W rather than 73°W the ISOPYCNIC WBC transport would increase to 14 Sv whereas the interior transport decreases to 6 Sv), the total transport of 20 Sv at 25°N in ISOPYCNIC is substantially above both other models (13 resp. 14 Sv).

It is also interesting to note that the overturning in LEVEL is generally larger than in the previous CME runs, even in exp. K13-6 which used a relaxation to an observed hydrographic section to represent the effect of overflow. Also, it is seen that the wind has, at least on time scales of a few years, little influence on deep boundary currents although it greatly affects the Western Boundary Current.



(a) LEVEL



(b) ISOPYCNIC

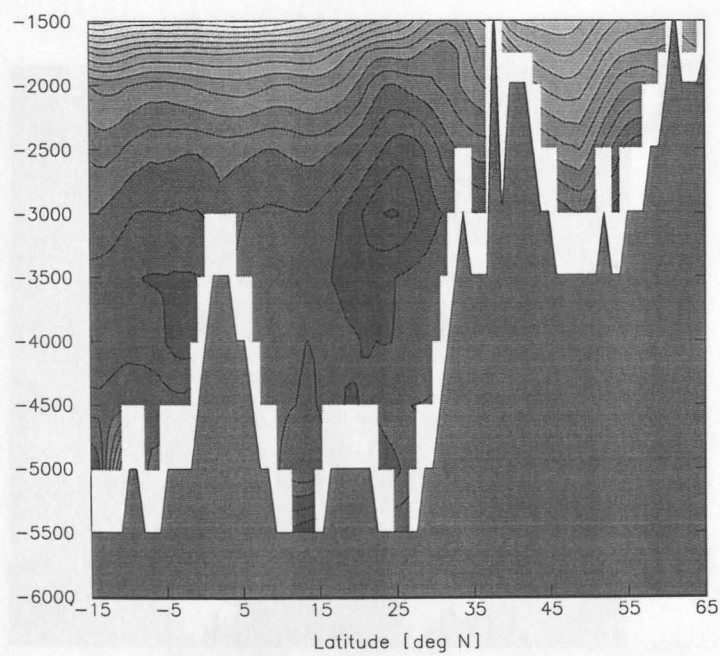
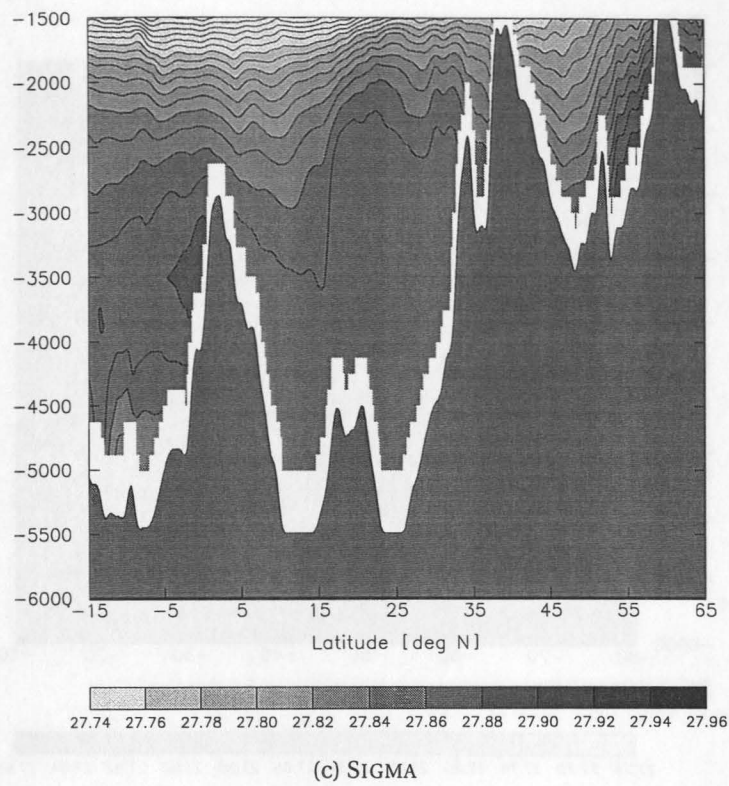
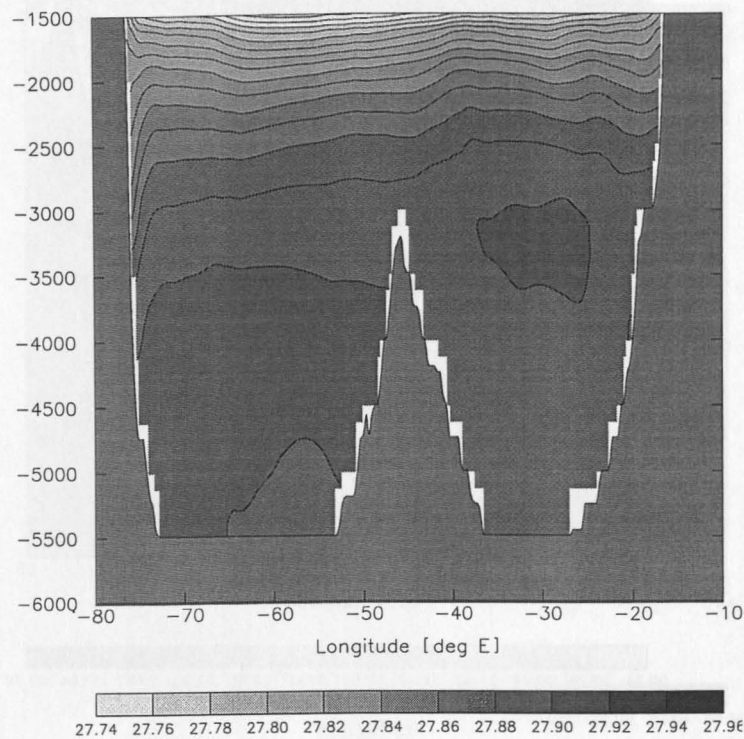
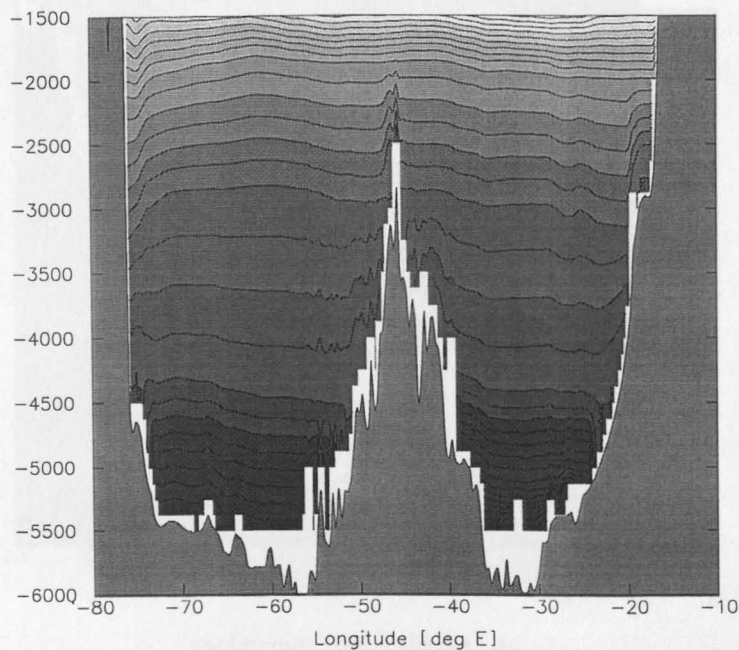


Figure 4.5: Density variable σ_θ along 30°W for all models, and for the LEVITUS (1982) climatology. Contour interval is 0.01 sigma-units.



(a) LEVEL



(b) ISOPYCNIC

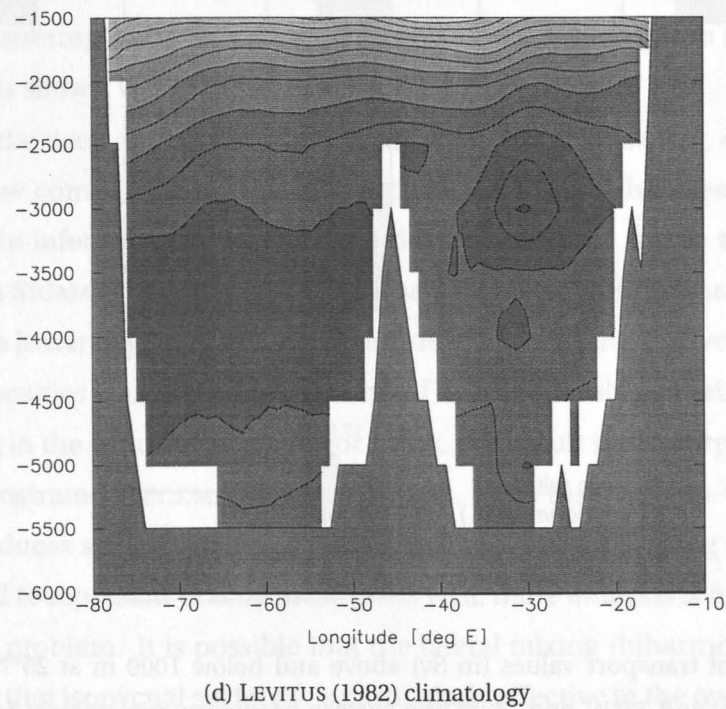
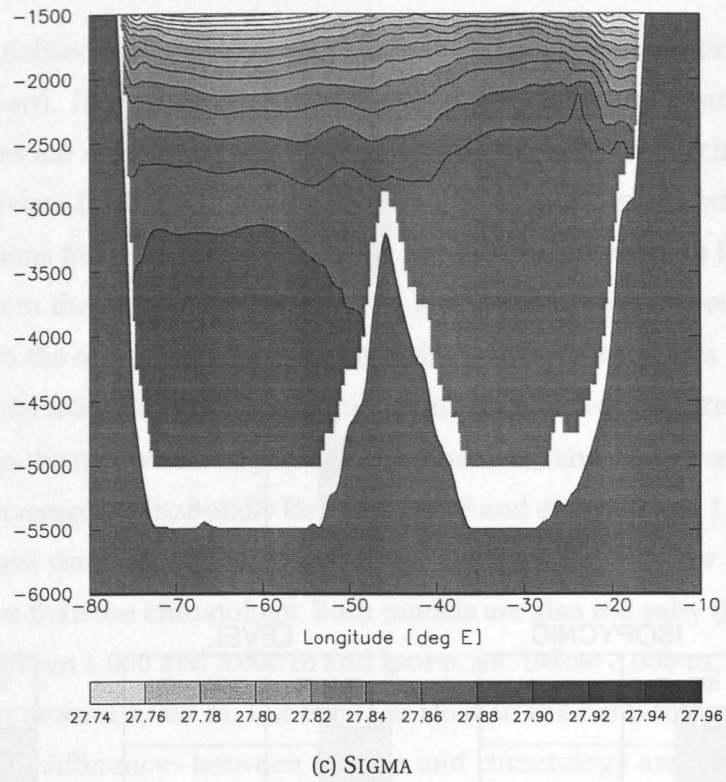


Figure 4.6: Density variable σ_θ along 25°N for all models, and for the LEVITUS (1982) climatology. Contour interval is 0.01 sigma-units.

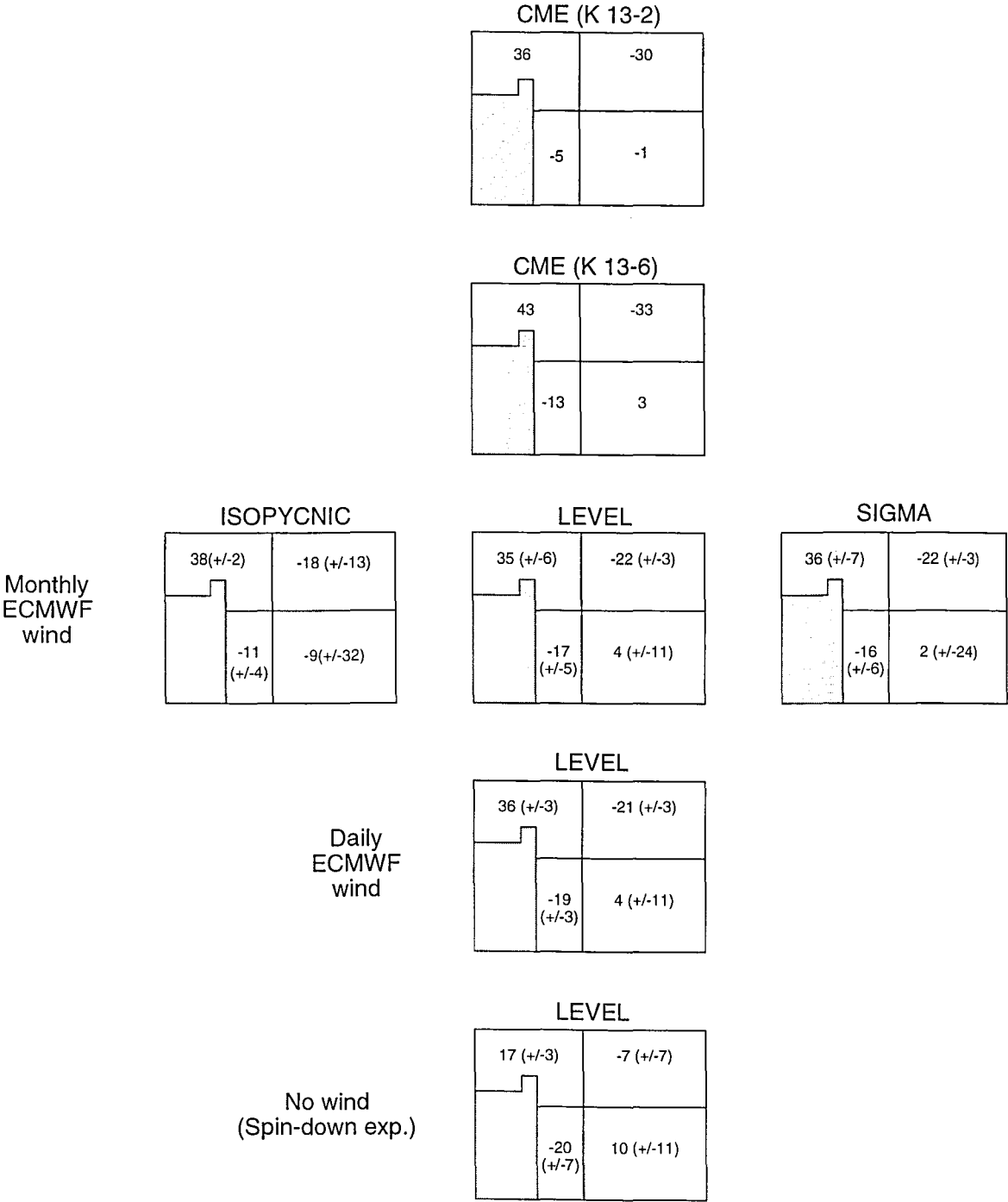


Figure 4.7: Partion of transport values (in Sv) above and below 1000 m at 25°N, between western boundary and 73°W, and 73°W and eastern boundary, respectively, for the three models. The first value gives the net transport, the second the amount of recirculation through the respective section.

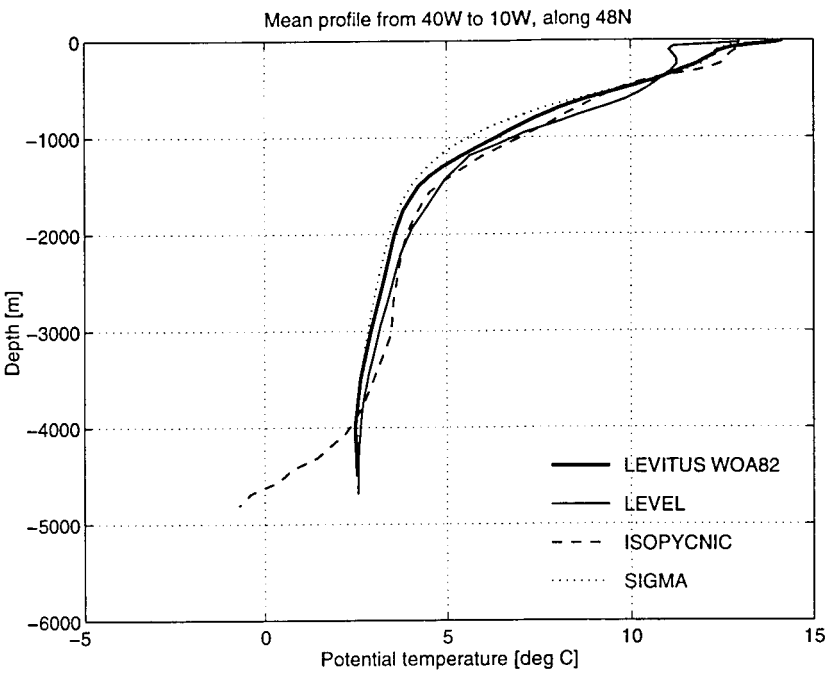
4.1.3 Water mass properties

All models are initialised with the LEVITUS (1982) climatology, except north of the ridges (cf. year 1 project report). In regions which have been in contact with the atmosphere, the water mass properties are at least partially determined by the ECMWF forcing and must hence be expected to deviate from the initial distribution. In deeper regions which have not been ventilated, deviations from the climatology may however be taken as an indication of model problems, at least to the extent that the LEVITUS atlas correctly describes the oceanic water mass relations. On the other hand, agreement with the climatology does not prove anything as the models are far from thermohaline equilibrium after 20 years of integration.

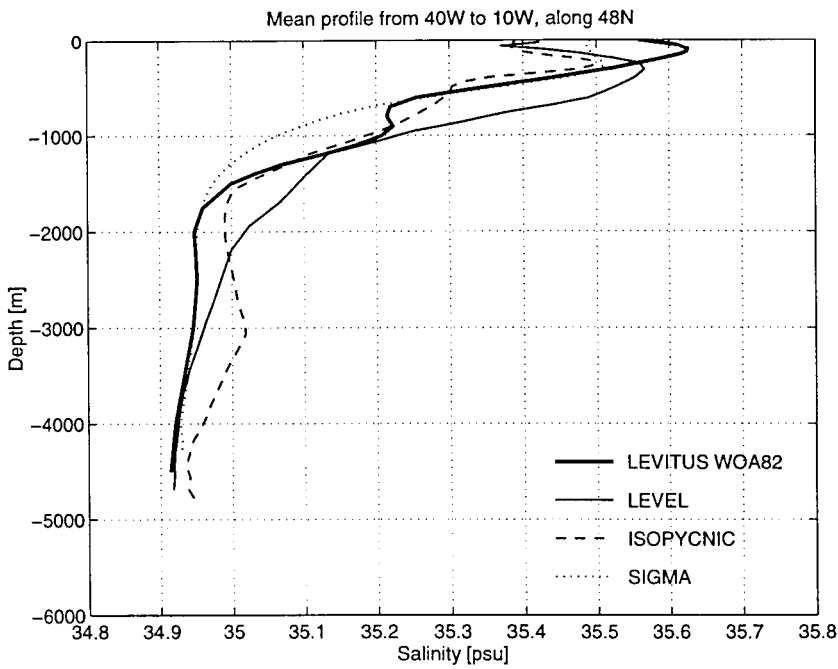
Fig. 4.8 displays the mean vertical profiles of temperature and salinity at 48°N in the north-eastern Atlantic, averaged horizontally between 10°W and 40°W. Below 1,000 m, both LEVEL and ISOPYCNIC have warmed slightly, by up to 0.5°C (Fig. 4.8(a)). Below 4,000 m ISOPYCNIC is up to 3 °C colder than the climatology. Both models are also too salty (Fig. 4.8(b)) by up to 0.1 psu, LEVEL between 1,000 and 3,000 m and ISOPYCNIC below 2,000 m. SIGMA has become cooler and fresher around 1,000 m, but is rather close to the climatology at greater depths. At 25°N Fig. 4.9 the differences between models and climatology are smaller, reflecting the larger distance from the source region. Fig. 4.10 shows the evolution of deep temperatures in the centre of the subtropical gyre. ISOPYCNAL cools by 0.2°C during the first 10 years, while both other models show a weak warming trend.

Examples for the deep density distribution are given along 30°W (Fig. 4.5) and along 25°N (Fig. 4.6). To allow comparison with ISOPYCNIC, the variable σ_θ has been chosen although that is not ideal to infer characteristics of the deep circulation. While the structure of the density section in SIGMA and LEVEL resembles the observed distributions, the deep densities in SIGMA and, to a lesser degree, also LEVEL are indeed lower than observed values. It appears that the higher densities which are present north of the sills vanish, probably due to too strong diapycnal mixing in the overflow regions. For LEVEL this result is not surprising, it has in fact often been demonstrated (BECKMANN and DÖSCHER, 1997) that the step-wise representation of topography induces strong diapycnal mixing in flows down a sloping bottom. For SIGMA which is designed to represent exactly those flows well, more analysis is needed to determine the origin of this problem. It is possible that the lateral mixing (biharmonic diffusion along horizontal rather than isopycnal surfaces) is particularly effective in the overflow region where the isopycnal slopes are steep.

The density distribution in ISOPYCNIC differs substantially from both the other models as well as from the observations. The deep densities significantly exceed the climatological val-

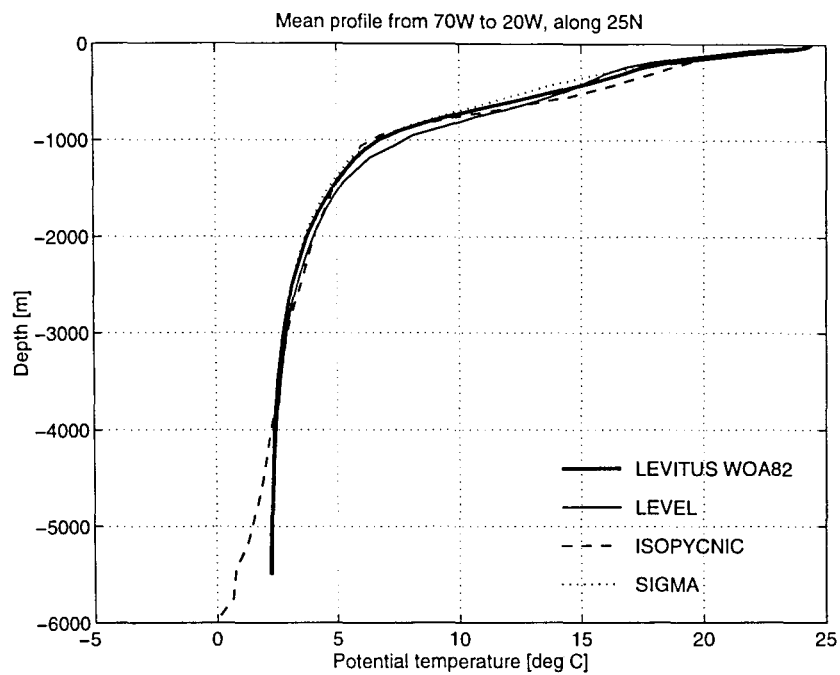


(a) Mean potential temperature.

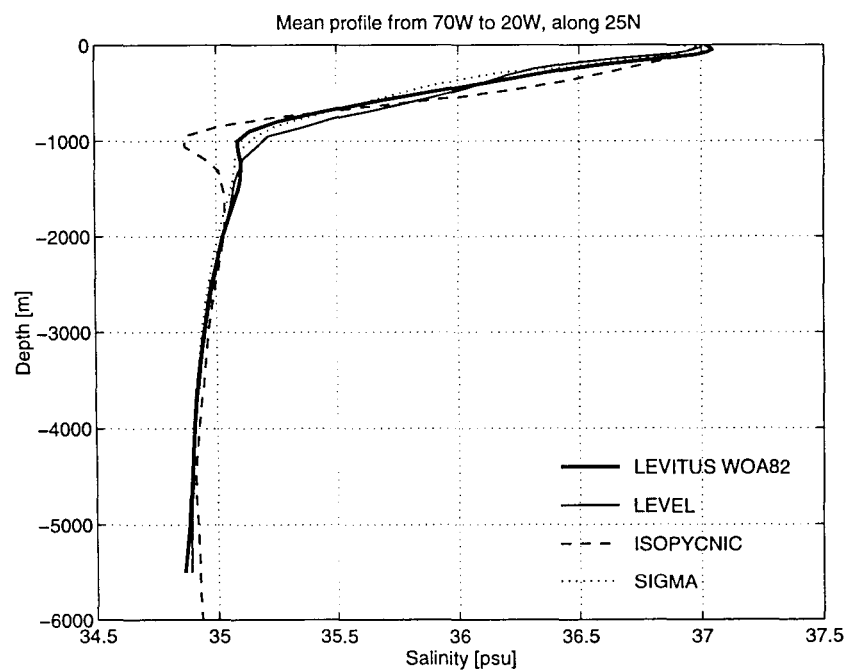


(b) Mean salinity.

Figure 4.8: Mean potential temperature and salinity vs. depth at 48°N, averaged between 10°W and 40°W.



(a) Mean potential temperature.



(b) Mean salinity.

Figure 4.9: Mean potential temperature and salinity vs. depth at 25°N, averaged between 20°W and 70°W.



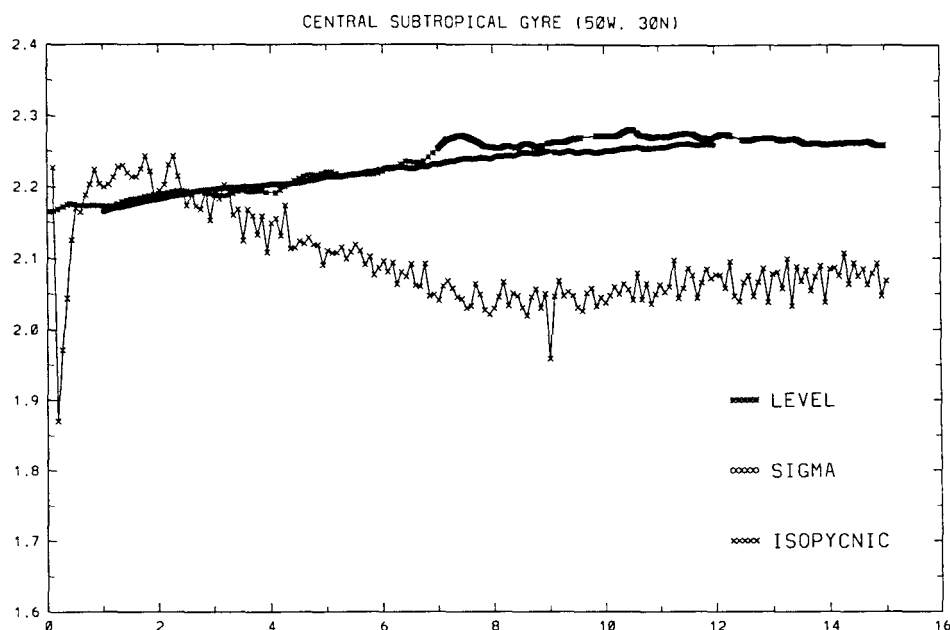


Figure 4.10: Evolution of deep temperature during the spin-up integration for the three models.

ues everywhere, in particular already in the subpolar region. It appears that in the overflow region there is more mixing in nature than in ISOPYCNIC. The diapycnal mixing in ISOPYCNIC is formulated through explicit diffusion, with a coefficient according to GARGETT (1984). It is a very rare situation for an ocean modeler to find that more, rather than less, mixing is needed. Increasing the mixing coefficient is not difficult, and would probably improve the situation, in particular if that increase were restricted to areas with strong topographic slopes where diapycnal mixing appears to be strongest (TOOLE et al., 1994). Also, the effective diapycnal mixing would be increased if the processes of thermobaricity and cabbeling which arise through the nonlinearity in the Equation of state were incorporated into ISOPYCNIC, although it is not clear what effect such increase would have on the thermohaline circulation as in particular the latter process induces non-diffusive diapycnal transports that tend to increase the density.

4.1.4 Meridional heat transport

The meridional heat transport is a variable of high climatological interest. While it is related to the zonally-averaged heat uptake, differences in heat storage of the models which are not in full thermal equilibrium may complicate the interpretation of this variable which should therefore be considered with some caution.

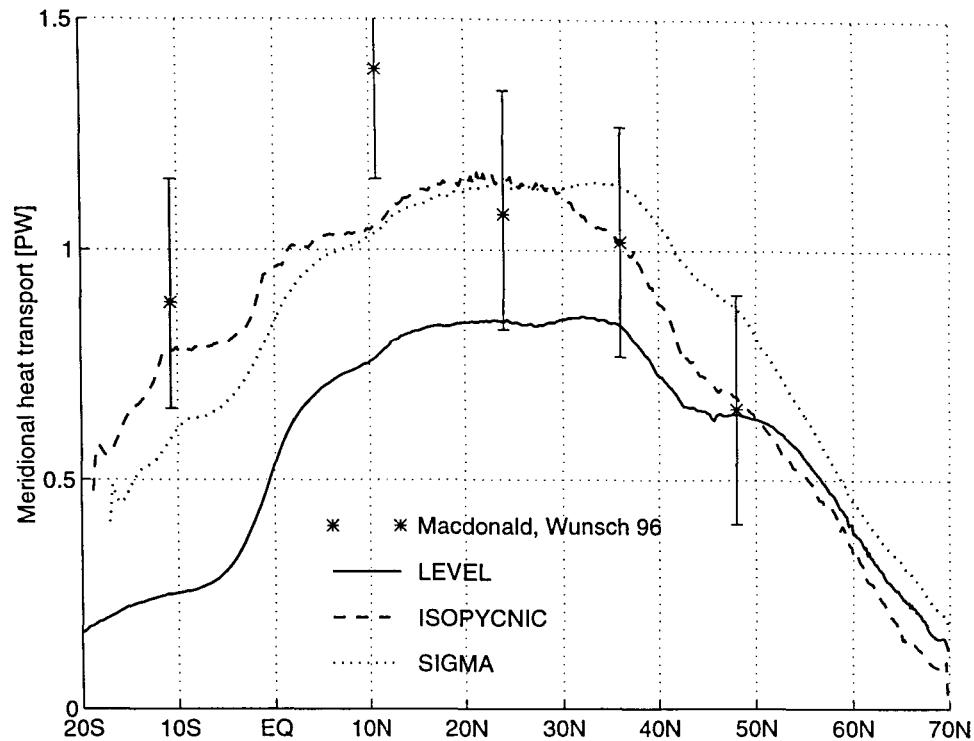


Figure 4.11: Northward transport of heat in all three models.

Some distinct differences between the three models are apparent in Fig. 4.11. Both ISOPYCNIC and SIGMA are fairly close in this variable, and reach a maximum of almost 1.2 PW near 20°N (ISOPYCNIC) and between 20–35°N (SIGMA). North of 35°N the SIGMA heat transport is significantly (up to 0.2 PW) larger. North of 20°N both models are within the error bars of the recent estimates from hydrographic sections (MACDONALD, 1996), and south of 20°N the models are somewhat lower. The heat transport in LEVEL is substantially (up to 0.4 PW) below both other models and the observations, except in the subpolar region, it reaches a maximum of only 0.85 PW between 20–35°N. Qualitatively, the difference between LEVEL and ISOPYCNIC is similar to the intercomparisons by CHASSIGNET et al. (1996). It is remarkable that all models are very close at 60°N, predicting a value of 0.4 PW which is on the high side of most published estimates.

The much too low heat transport in LEVEL has various causes. The overall strength of the meridional transport is weakest in LEVEL, for the reasons discussed above. As the relation between heat transport and overturning rate at 25°N for all models is roughly similar to that of a number of integrations discussed by BÖNING et al. (1996) (cf. Fig. 4.12), it is likely that the weaker overturning is mainly responsible for the lower heat transport in LEVEL. A second im-

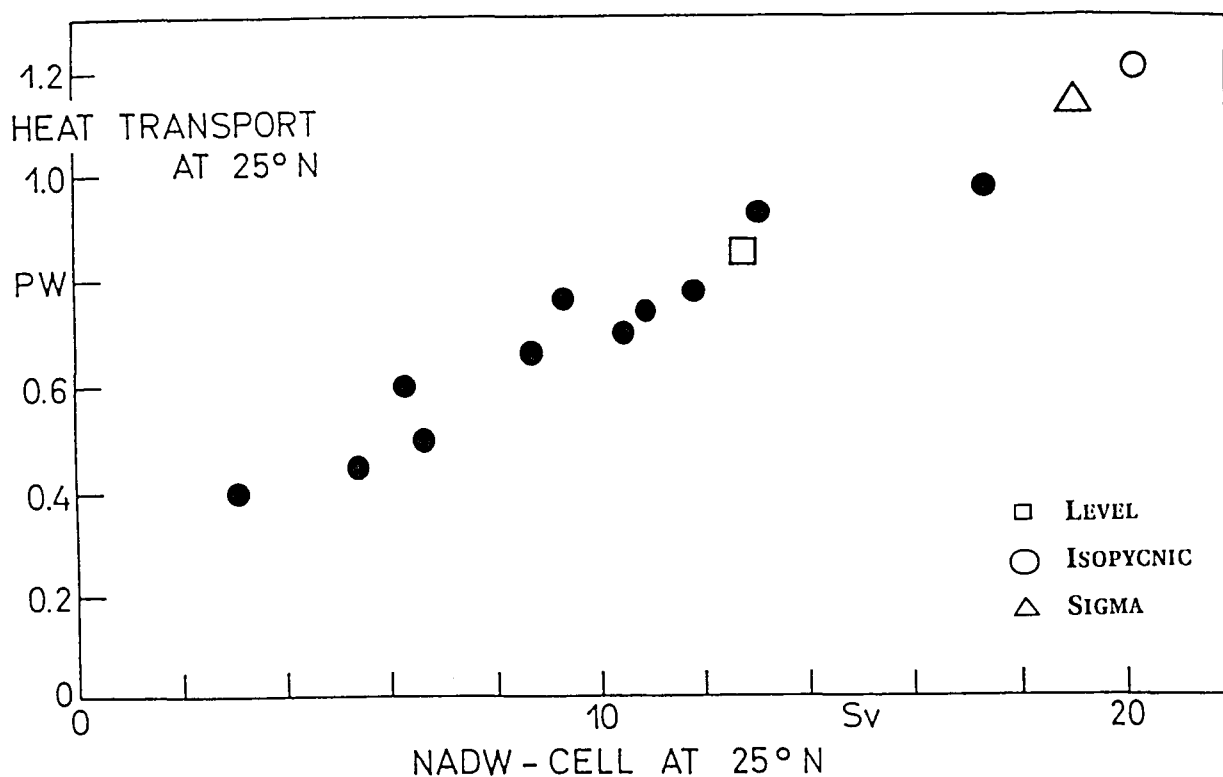
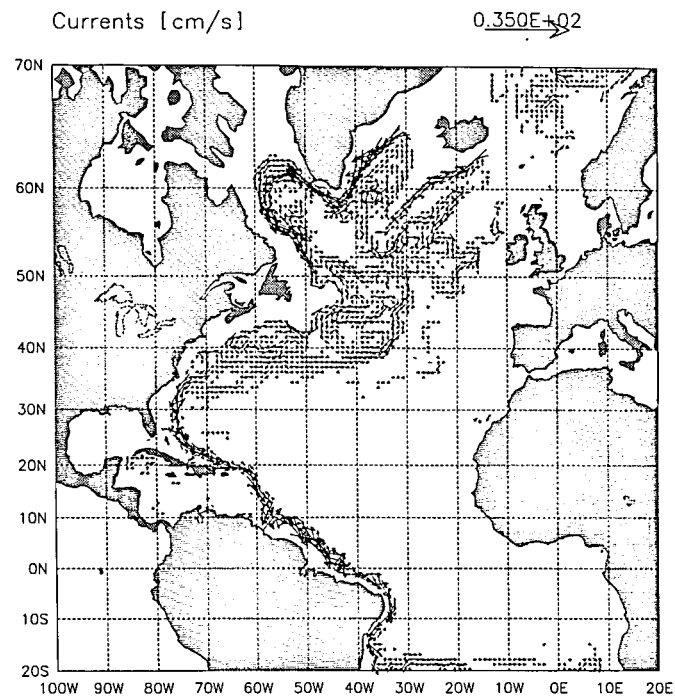
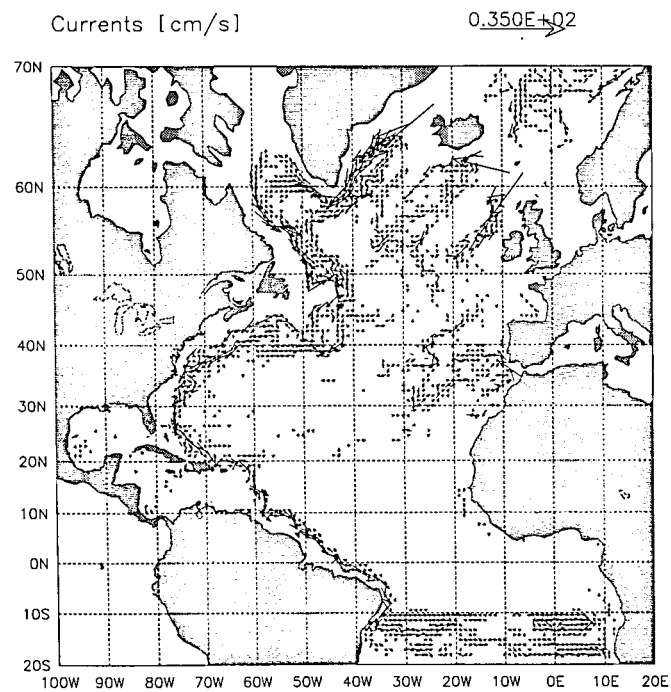


Figure 4.12: Heat transport vs overturning at 25°N for all three models, together with a number of other model integrations discussed by BÖNING et al. (1996).

portant factor is the formulation of the southern boundary condition which differs in LEVEL (open boundary) from the other two models (closed boundary with restoring zone). A sensitivity experiment with LEVEL where inadvertently the sign of the (prescribed) barotropic flow at the southern boundary was altered, shows a dramatic change in overall LEVEL heat transport, with a maximum reaching 1.15 PW and the value at the southern boundary increased from 0.2 to 0.7 PW. In the region of the North Atlantic Current between 40°N and 50°N, where the LEVEL heat transport divergence is less than 0.1 PW as compared to nearly 0.3 PW in both other models, the LEVEL circulation differs from the other models as will be discussed below.



(a) LEVEL



(b) ISOPYCNIC

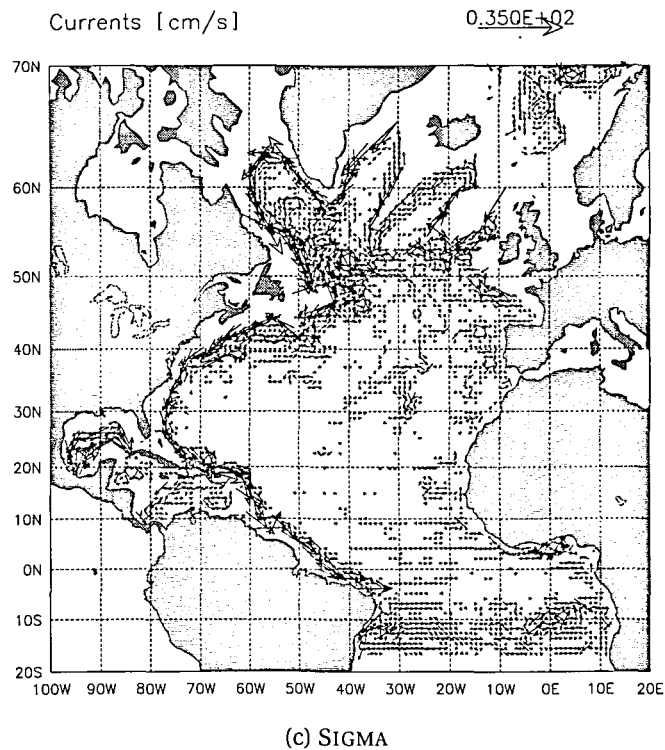


Figure 4.13: Horizontal velocity of all models at 1625 m depth.

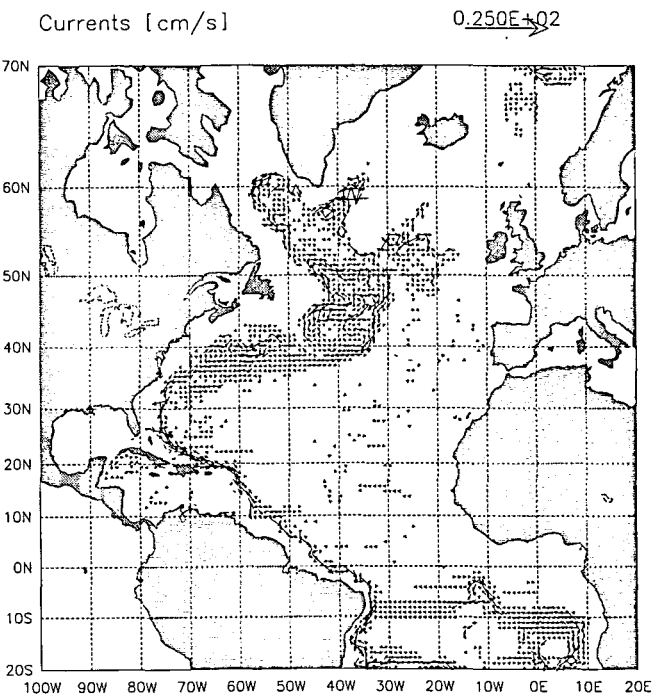
4.1.5 Deep circulation

The circulation at 1,600 m depth for all three models is displayed in Fig. 4.13. South of 30°N, all models show a rather similar pattern for the Deep Western Boundary Current. The DWBC transport in ISOPYCNIC is weakest, as to be expected from the vertical distribution of transports discussed above. Further north, there are however considerable differences between the models. In SIGMA and especially in ISOPYCNIC the circulation is confined to the western boundary up to 50°N, whereas LEVEL has a southward transport along the western flank of the Mid-Atlantic Ridge between 40 and 50°N, with evidence for a northward recirculation south-east of the Grand Banks. That transport appears to be connected to the different NAC pathway in LEVEL which flows north also at 30°W (cf. section 4.3 below). Also, in LEVEL the westward flow below the Gulf Stream is much broader and appears to be split into two branches. The circulation in the subpolar North Atlantic will be described in more detail in chapter 5 below.

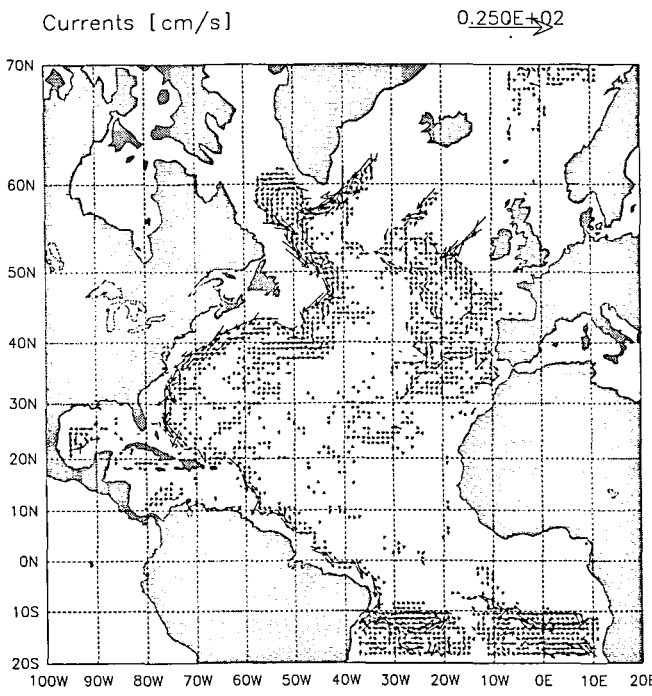
In the eastern basin, all models show rather weak circulation, with southward flow between 20 and 30°W which generally is largest in SIGMA. In ISOPYCNIC a small cyclonic cell

off Gibraltar is visible which is associated to a penetration of Mediterranean Water to deeper levels.

Most of the patterns found at 1,600 m depth also dominate the circulation at 2,800 m (Fig. 4.14). More pronounced are differences between the models in the eastern basin where LEVEL has a very weak circulation. In ISOPYCNIC and to some degree also in SIGMA, water resulting from the overflow through the Faroe-Scotland Channel appears to be transported in a boundary current along the eastern flank of the Mid-Atlantic Ridge (MAR). SIGMA has an anticyclonic gyre in the subtropical eastern basin which is related to the barotropical circulation component (cf. Fig. 4.15 below). In the tropics, the circulation in SIGMA is substantially stronger than in both other models.



(a) LEVEL



(b) ISOPYCNIC

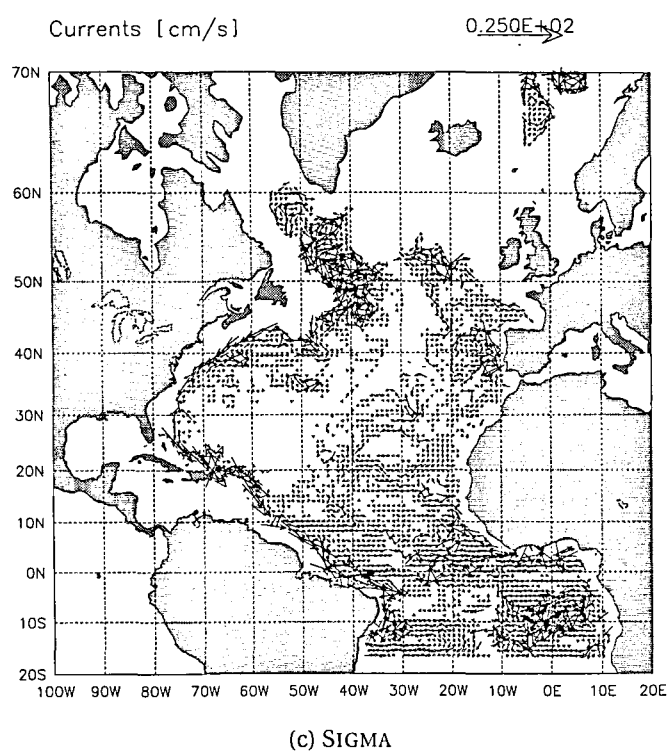
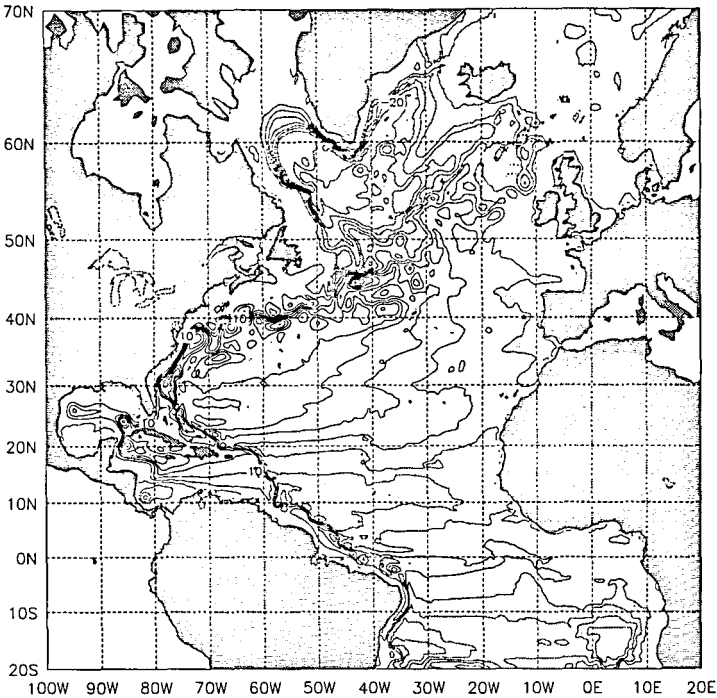
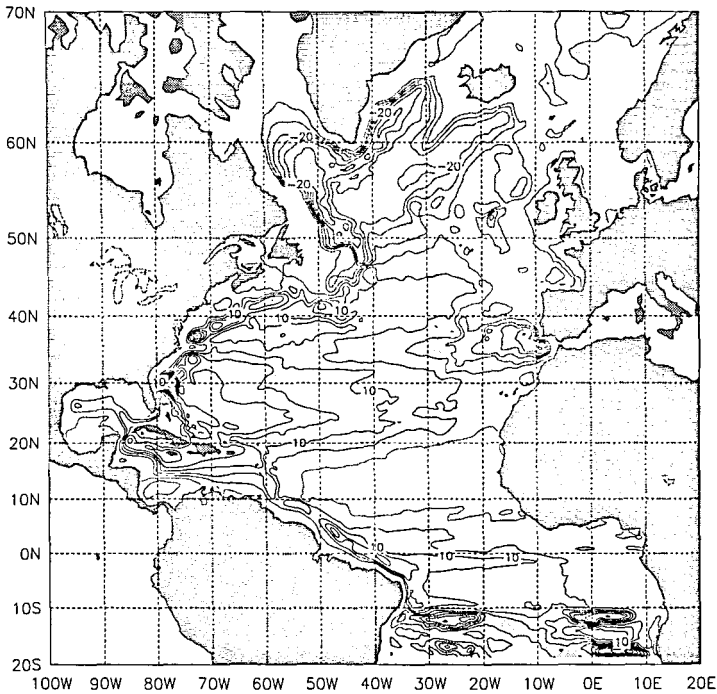


Figure 4.14: Horizontal velocity of all models at 2875 m depth.



(a) LEVEL



(b) ISOPYCNIC

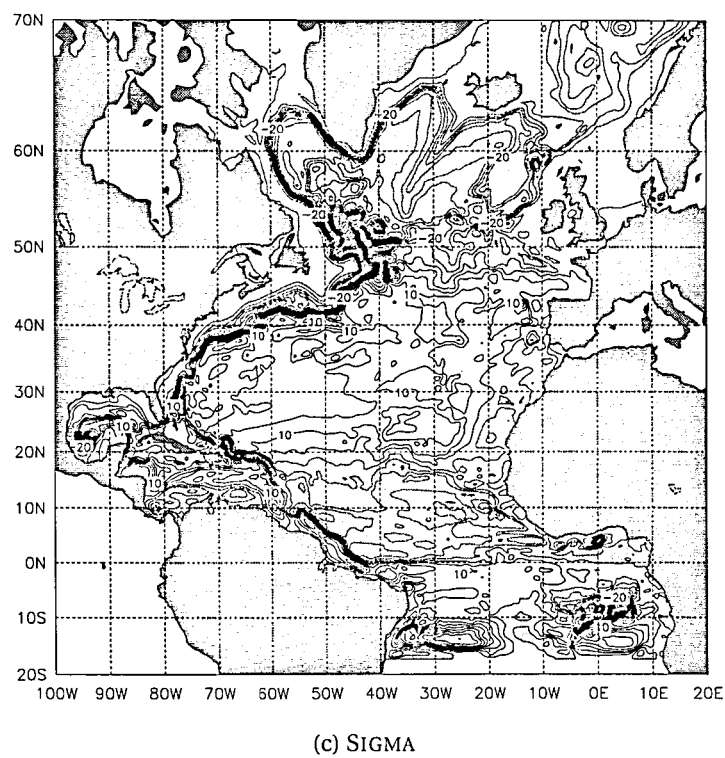


Figure 4.15: Streamfunction of vertically integrated mass transport for all models.

4.2 Aspects of the wind-driven circulation

4.2.1 Barotropic transport

The vertically integrated transport is a frequently-used variable to describe the overall horizontal circulation pattern. While often an expression of the circulation in the upper kilometre, the interpretation of the vertically integrated transport may however not be very straightforward as it constitutes a sum over flows that may have quite different origin and dynamics. The 5-year mean barotropic streamfunction for the three models which is shown in Fig. 4.15 has a rich structure, with much of the transport confined to rather narrow regions near the continental shelf break. All models have well-developed gyres of roughly similar structure and magnitude, reaching approximately 30 Sv in the subpolar gyre south of Cape Farewell, and 30–35 Sv in the Gulf Stream at 25°N. All models show some narrow recirculation regimes associated with the WBC, which differ however in detail. The recirculation regimes are most pronounced in SIGMA which in particular displays a series of rather strong transport cells northeast of the Grand Banks. Although the barotropic circulation component usually is not well observed, there is some evidence that those structures might be not unrealistic (ROSSBY, 1996).

Two closed recirculation cells are found in the eastern basin which appear to be somewhat unrealistic. SIGMA displays a cyclonic cell in the eastern basin, reaching 10 Sv with the center at 25°N. At this latitude, one would expect the Sverdrup balance to hold to a good first approximation. As seen in Fig. 4.16, this appears by and large to be the case for LEVEL and ISOPYCNIC which agree well east of 55°W. In SIGMA, however, a strong deviation from Sverdrup balance associated to the anticyclonic gyre is visible in the eastern basin. ISOPYCNIC (Fig. 4.15(b)) shows a rather strong anticyclonic cell off Gibraltar with a magnitude exceeding 20 Sv. This feature is likewise inconsistent with a Sverdrup balance which at 36°N one would expect at least in the eastern basin. It is probably related to the “Azores Current” which occurs at this latitude in ISOPYCNAL but not in the other models (cf. chapter 7).

On smaller scales, the models are less consistent with each other. In the Gulf Stream and North Atlantic Current regions, numerous closed recirculation scales are found which are by far strongest in SIGMA, weaker in LEVEL but almost absent in ISOPYCNIC. As discussed in chapter 8, these are the regions of highest eddy activity in all three models. As the eddy activity in ISOPYCNIC is much lower than in the other models, it appears likely that these small-scale cells are directly caused by the eddies. It is unclear however whether the cells are attached to small-scale topographic patterns, or whether averaging over a longer period

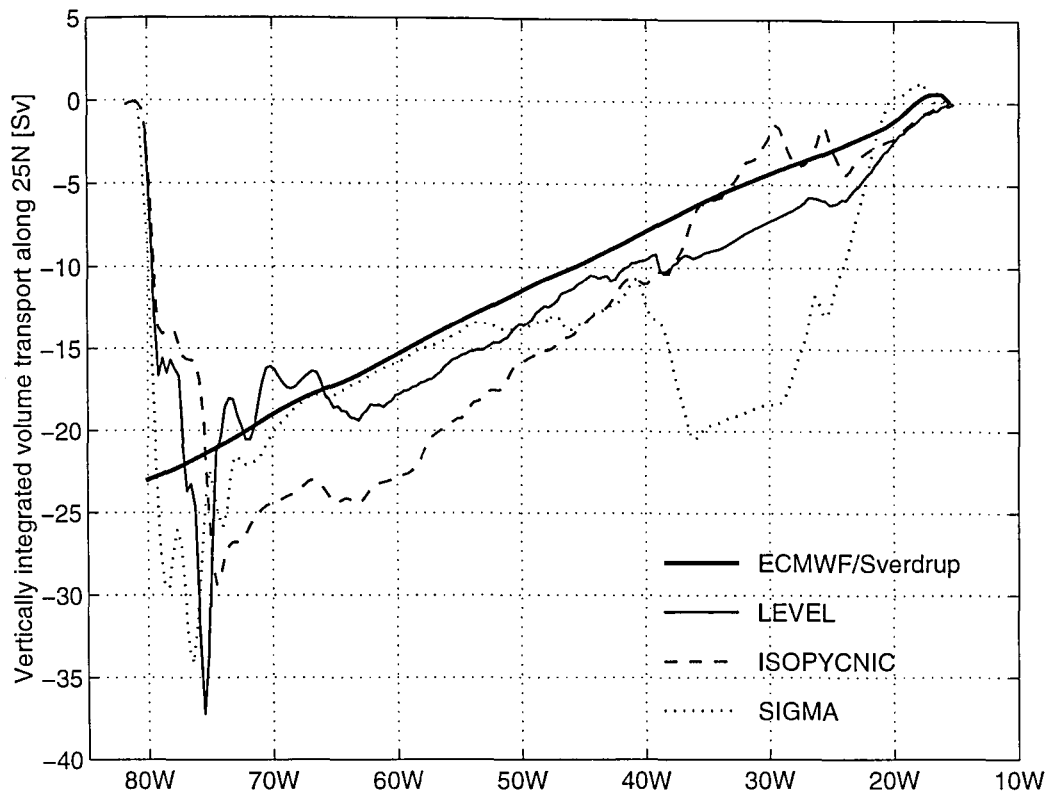


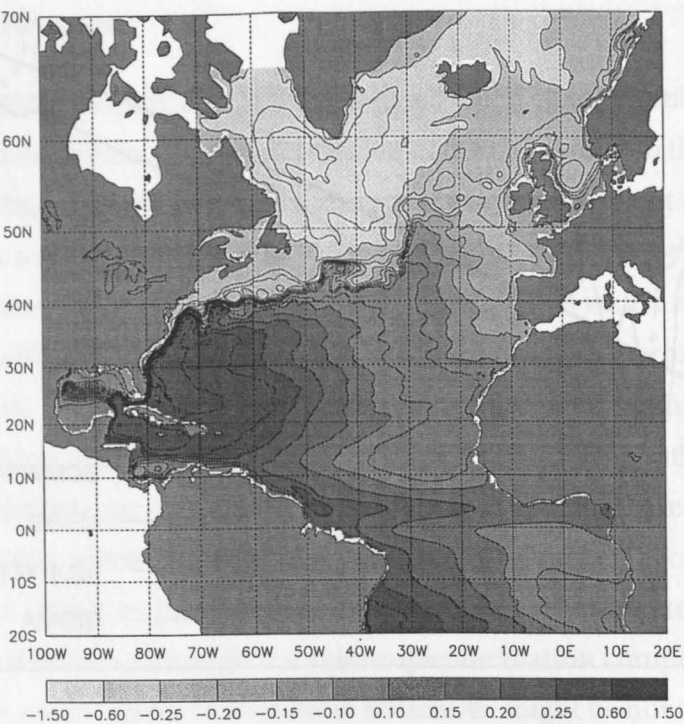
Figure 4.16: Barotropic transport for all models at 25°N.

would reduce their amplitude.

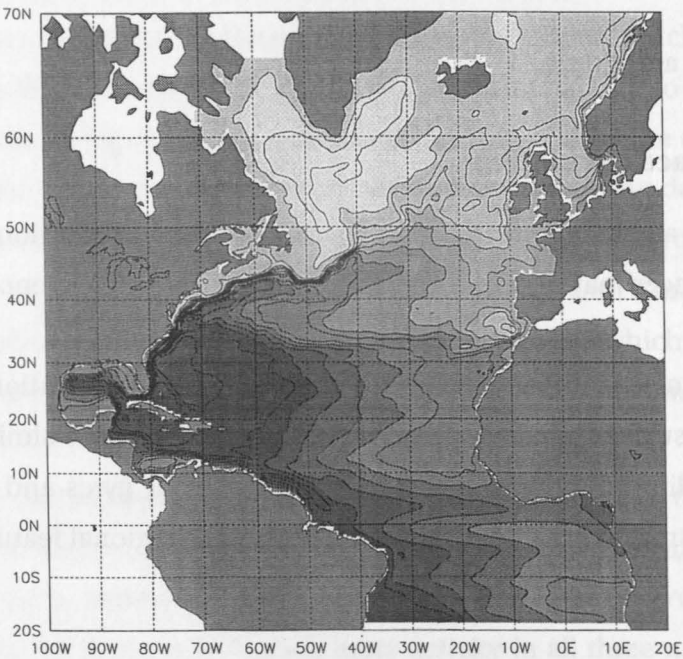
4.2.2 Near-surface circulation

In the upper ocean all models are in dynamical equilibrium. In addition to the vertical structure, model differences may also be the result of the inevitable differences in the mixed-layer formulation.

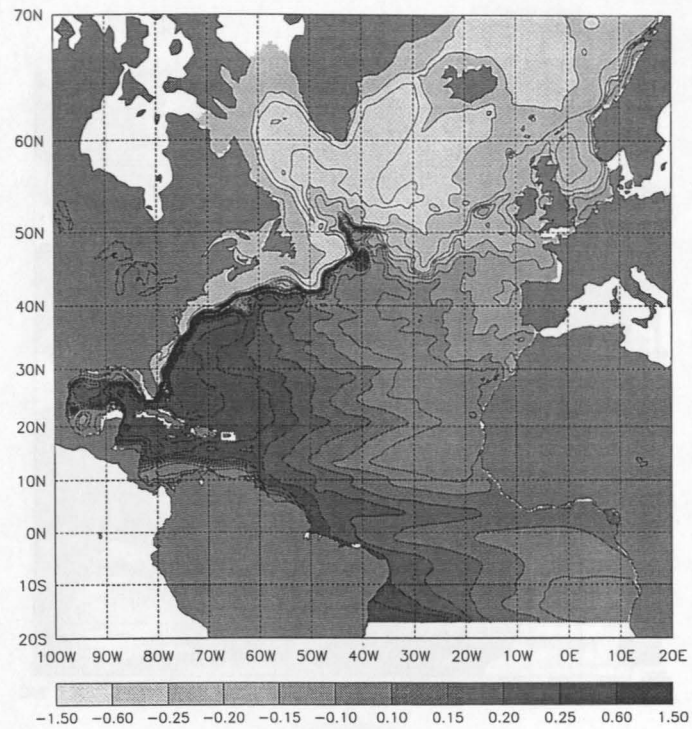
The 5-year mean sea-surface elevation (Fig. 4.17) gives an indication for the geostrophic component of the surface circulation. All three models are rather similar in the large-scale structure and amplitudes, both in the subtropical/subpolar gyres and in the equatorial region, but differ on smaller scales and in several important regional features.



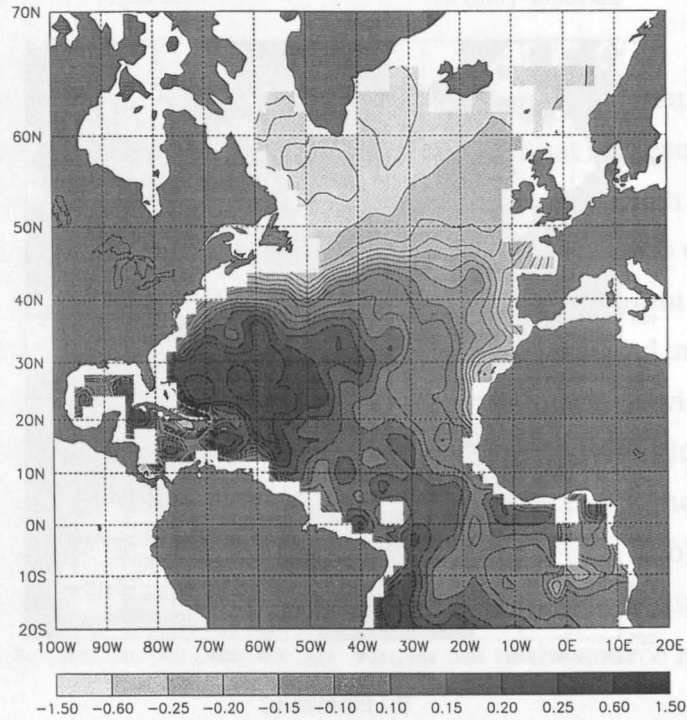
(a) LEVEL



(b) ISOPYCNIC

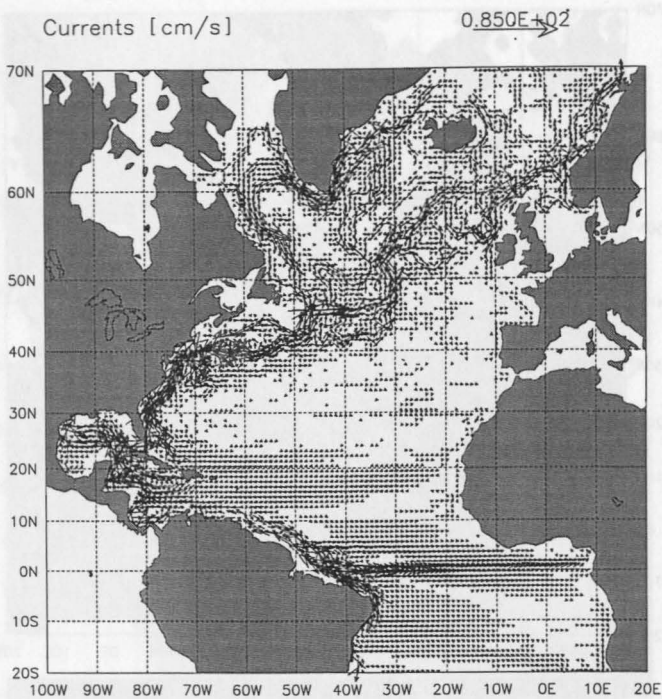


(c) SIGMA

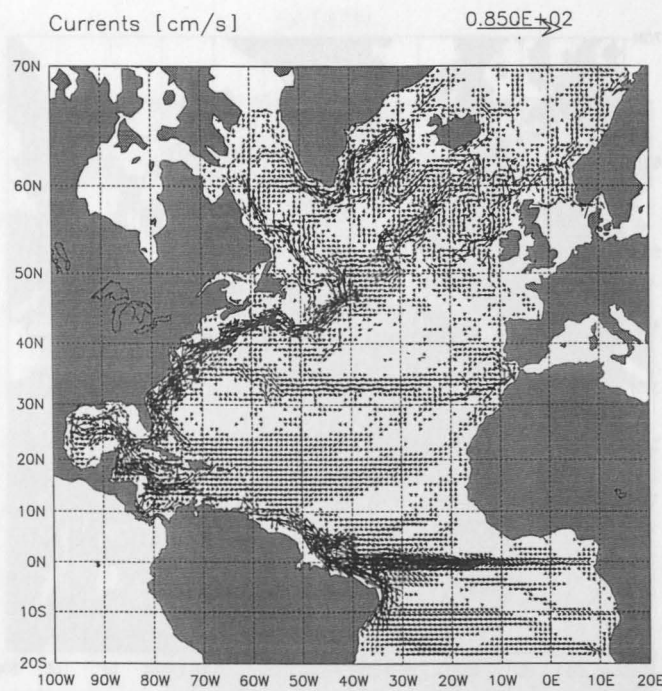


(d) TOPEX/Poseidon

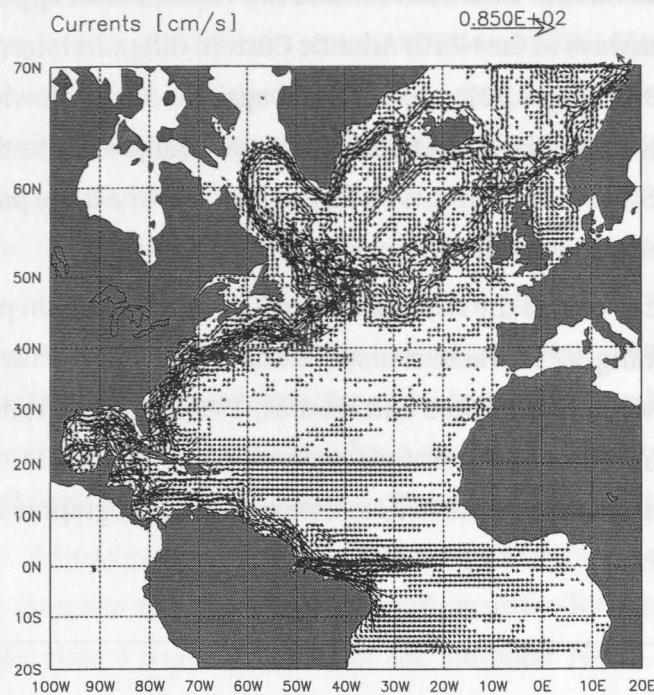
Figure 4.17: Mean sea-surface elevation for all three models and TOPEX/Poseidon.



(a) LEVEL



(b) ISOPYCNIC



(c) SIGMA

Figure 4.18: Horizontal velocity of all models at 92 m depth.

None of the three models is able to simulate the observed separation of the Gulf Stream at Cape Hatteras. This is not an unexpected result, it is likely that the resolution in all models is still too coarse to capture the details of the vorticity dynamics which control the separation process (DENG et al., 1996). Downstream, a strong recirculation is visible in ISOPYCNIC (Fig. 4.17(b)) and to some extent also in SIGMA (Fig. 4.17(c)) but absent in LEVEL which instead shows an anticyclonic recirculation in the Mid-Atlantic Bight and indications for standing eddies (Fig. 4.17(a)), a pattern that has also been observed in previous high-resolution simulations with the GFDL-model (BECKMANN et al., 1994). In ISOPYCNIC one branch of that recirculation turns eastward, forming an eastward current that crosses the basin at 33°N. That current is particularly visible in the current field at 92m depth (Fig. 4.18(b)), and has some resemblance to the Azores Current, albeit it occurs at a somewhat more southern latitude. Unlike the real Azores Current, it is however not connected to the source region of the NAC at Newfoundland.

A current with similar characteristics appeared in the simulation by BECKMANN et al. (1994). The eastward flow in the other two models is much weaker, and less coherent. A further discussion on the dynamics of this current system is given in chapter 7.

The flow of the Gulf Stream extension around the Grand Banks appears fairly realistic in all models, but the pathways of the North Atlantic Current differ. In ISOPYCNIC and SIGMA the NAC turns north at 42°W (Fig. 4.18), and eastward again at $52\text{--}54^{\circ}\text{N}$, with a branch towards the northwest which is particularly strong in SIGMA, reaching well into the Labrador Sea before turning back. In SIGMA, the currents in that region are for a large part eddy driven, with warm eddies generated off Flemish Cap.

In LEVEL only a small part of the NAC turns north at 42°W , the main part of the NAC flows more zonal until reaching 30°W where it finally turns north. This behaviour clearly deviates from the observed NAC pathway (KÄSE and KRAUSS, 1996). Results from earlier CME experiments, as well as from other high-resolution simulations with MOM (e.g. STAMMER et al., 1996) indicate a high sensitivity of the NAC to details of the topographic structure near Flemish Cap and Flemish Pass.

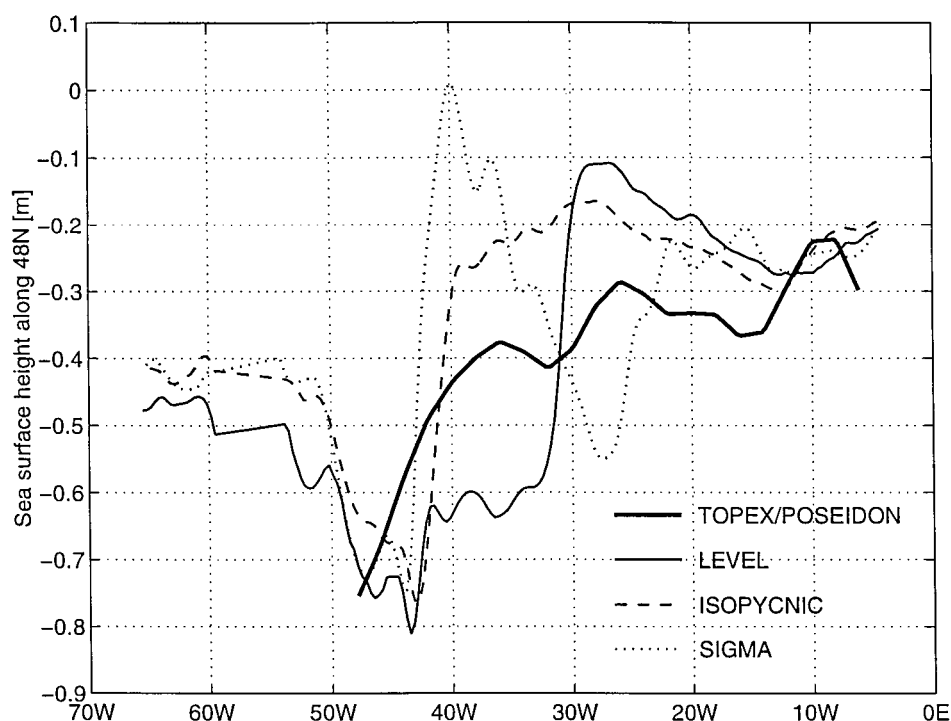


Figure 4.19: Mean surface elevation at 48°N for all models and from the TOPEX/POSEIDON altimetry.

The different pathways of the NAC are most clearly seen in the sea surface height at 48°N (Fig. 4.19). SIGMA has a steep SSH gradient of nearly 80 cm over 200km, indicating very strong northward flow at 42°W , but southward flow between 40°W and 30°W . The SSH-difference in ISOPYCNIC is 50 cm at roughly the same longitude but the flow between 40°W and 30°W re-

mains northward. LEVEL has its main gradient (50 cm) near 30°W. The gradient from the altimetry is less steep, amounting to 30 cm between 47°W and 37°W. However, one has to keep in mind that the altimetric estimate is necessarily smoothed, and also may contain some spurious signal from geoid undulations. All three models also have a fairly similar SSH decrease around 50°W associated to the Labrador Current.

Another common feature of all models is a substantial near-surface flow into the Norwegian Sea which continues into the Norwegian Current, in particular in SIGMA and LEVEL (Fig. 4.18). For the latter model, this constitutes a significant improvement compared to earlier CME results. In previous CME model studies with closed northern boundaries along 65°N and no artificial “hand-tuning” of the topography in the Faeroe-Shetland Channel the path of the NAC is restricted to the western North Atlantic flowing north into the Irminger Sea west of the Reykjanes Ridge. Allowing an throughflow across open boundaries along 65°N between Iceland and Norway does not alter the circulation significantly (REDLER and BÖNING, 1997).

Sensitivity studies with a regional model of the subpolar North Atlantic (REDLER and BÖNING, 1997) show that a northeastward flow of the NAC across the MAR toward the Rockall Plateau essentially depends on the realisation of the water exchange across the Iceland-Scotland Ridge. If the exchange is blocked due to an insufficient lateral resolution of the major deep pathway, the Faeroe Bank Channel, there is only a weak flow of upper layer water toward the Rockall-Faeroer region. Allowing an throughflow of cold Norwegian Sea Water at depths around 800 m into the eastern North Atlantic does not only alter the water mass characteristics in the deeper Iceland Basin but also affects the pathway of the NAC leading to the northeasterly flow feeding the Norwegian current.

4.3 Upper ocean

4.3.1 Ocean-atmosphere fluxes

The zonally averaged oceanic uptake of heat (Fig. 4.20) is fairly similar in all three models. The equatorial uptake of heat is strongest in LEVEL, and somewhat weaker in ISOPYCNIC. In the subtropical gyre between 10°N and 30°N, all models have a nearly vanishing net flux that is within the error bars of the recent climatological estimate by HASSE et al. (1996) but well above the average heat flux diagnosed from the ECMWF forcing. This is a remarkable result as the ECMWF data were used as forcing, and implies that the surface temperature of all models must be systematically colder than the SST from the forcing field. This is borne out in Fig. 4.21 which shows that the zonally averaged surface temperatures of all models are fairly close but

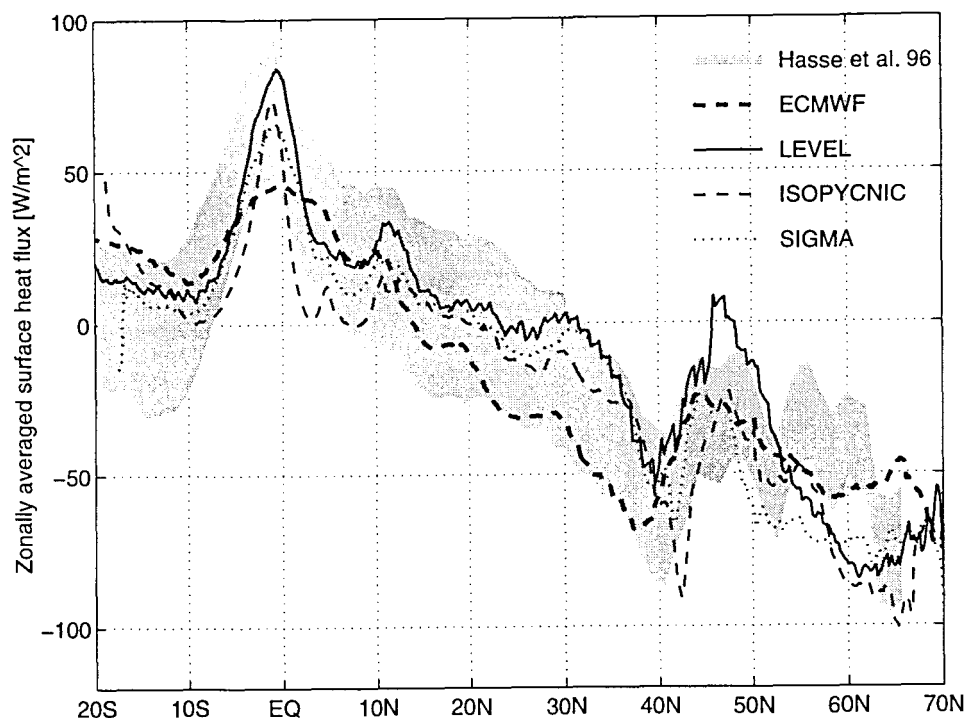


Figure 4.20: Zonally averaged surface heat flux in W/m^2 vs. latitude for all three models, and for the ECMWF forcing. Positive when ocean gains heat. The shaded area corresponds to the climatological estimate by HASSE et al. (1996).

systematically $1\text{--}2^\circ\text{C}$ colder than the ECMWF forcing temperature.

Near 40°N all models lose a substantial amount of heat, from 50 W/m^2 in LEVEL to 90 W/m^2 in ISOPYCNIC. Near 45°N , the heat loss is substantially reduced, and in LEVEL even turned into a weak gain. The quantitative differences between the models are larger here, reflecting the different structure of the North Atlantic Current system in the models (see below). Around 60°N the models are again close to each other and feature a zonally averaged heat loss of $50\text{--}70 \text{ W/m}^2$, in contrast both to the climatological as well as the ECMWF data, implying that all are systematically warmer than the forcing SST.

The horizontal distribution of ocean-atmosphere energy flux (Fig. 4.22) is likewise rather similar in all three models, and differs distinctly from the ECMWF forcing over much of the subtropical gyre where none of the models is able to accommodate a net heat loss through the surface (the net heat exchange is less than 10 W/m^2 in magnitude) whereas the forcing field expects an oceanic heat loss of $10\text{--}50 \text{ W/m}^2$. A heat loss of 25 W/m^2 would correspond to a cooling trend of the upper 100 m of more than 10°C over 5 years. It hence appears that in this region the model estimates are more plausible than the ECMWF-climatology for 1986–89

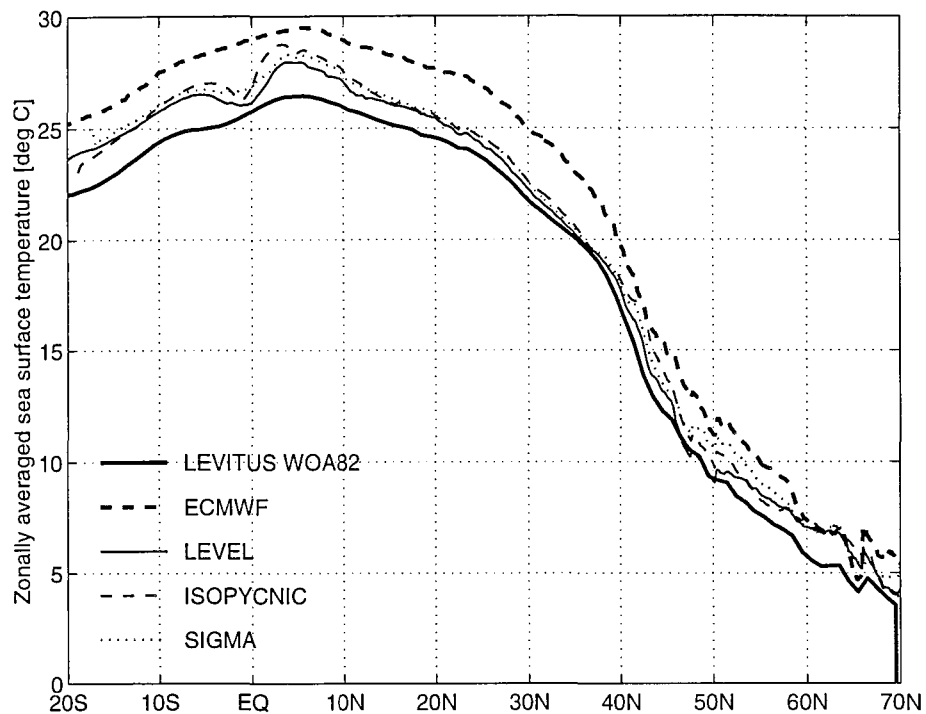
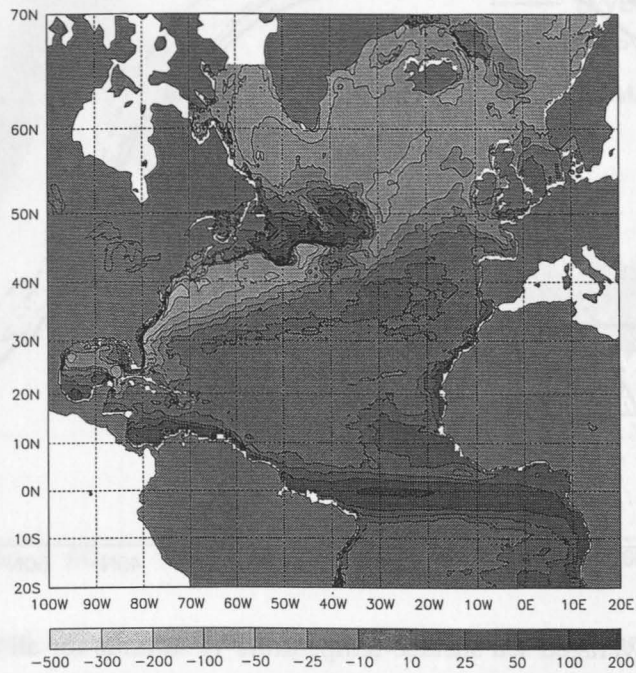


Figure 4.21: Zonally averaged sea surface temperature vs. latitude for all three models and for the ECMWF forcing.

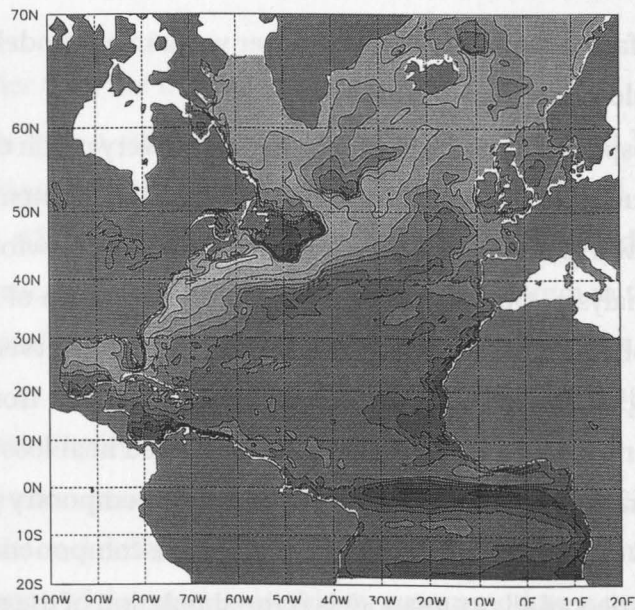
unless the near-surface advection which is rather weak in all models over most of the subtropical gyre (see below) were substantially too weak.

The ocean-atmosphere exchange near the Equator is very much concentrated in all models, reflecting the much higher resolution of the ocean models compared to the ECMWF analysis. All models show a small region of heat gain southeast of Newfoundland, in accordance with the ECMWF analysis. Only in LEVEL a secondary maximum of heat gain exists at 50°N and 30–40°W, a probably unrealistic feature that has been observed in earlier integrations with this model (e.g. SARMIENTO, 1986) and is due to advection from the northwest that is absent in both other models (cf. Fig. 4.18(a)). The reduced heat loss of the analysis near the Greenland and Labrador coastline probably is related to temporary presence of sea ice, and absent in the model estimates due to the lack of a sea-ice component.

A problematic aspect of all ocean models is the simulation of the net fresh water exchange with the atmosphere. All previous prognostic simulations with a boundary condition restoring to observed surface salinities have resulted in freshwater fluxes which at least regionally exceed a plausible range of magnitudes, including the inversion by SCHILLER (1995) which used the flux estimate by SCHMITT et al. (1989) as a constraint (in the least-squares sense).



(a) LEVEL



(b) ISOPYCNIC

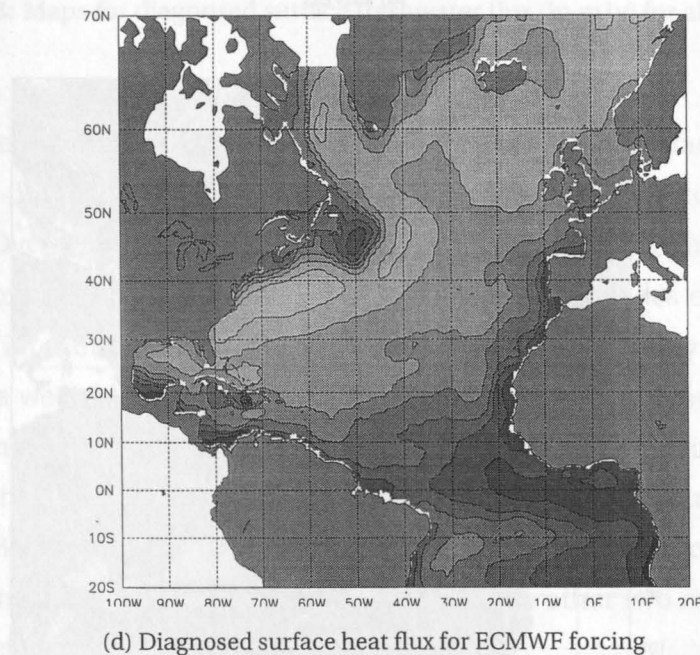
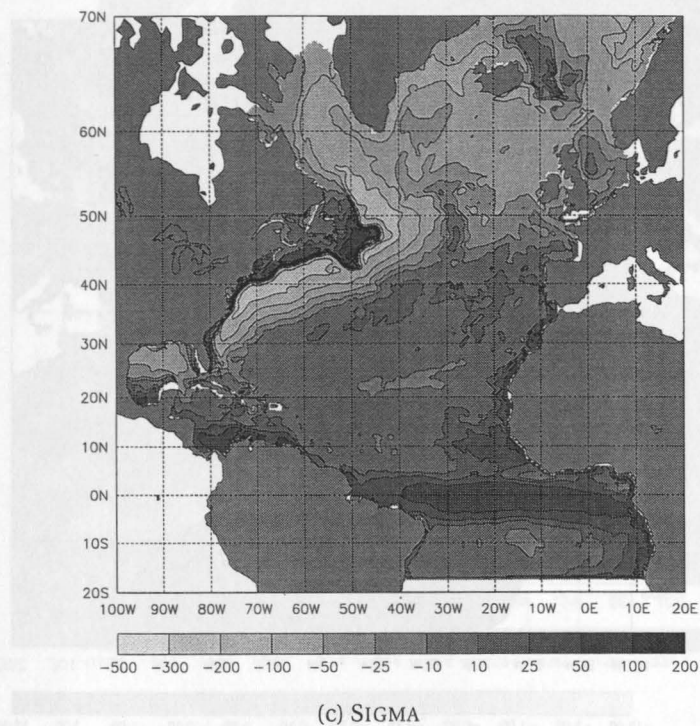
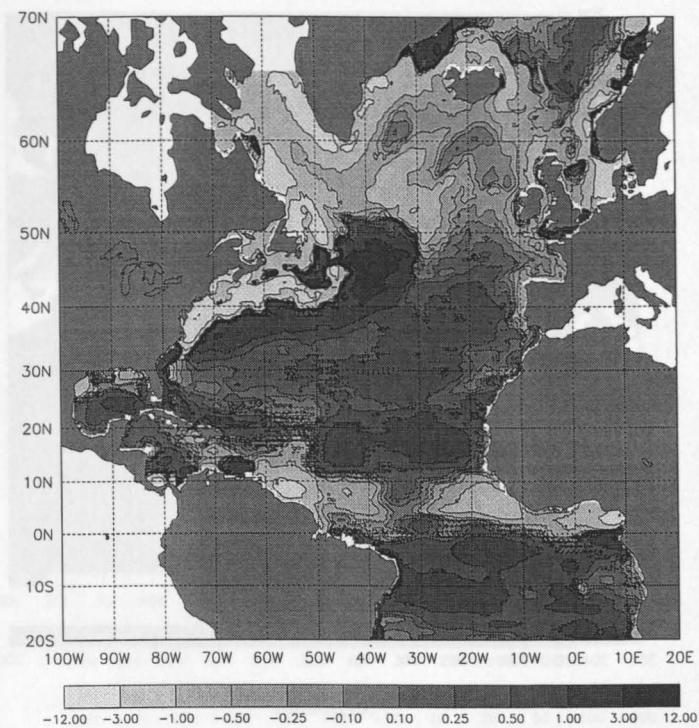
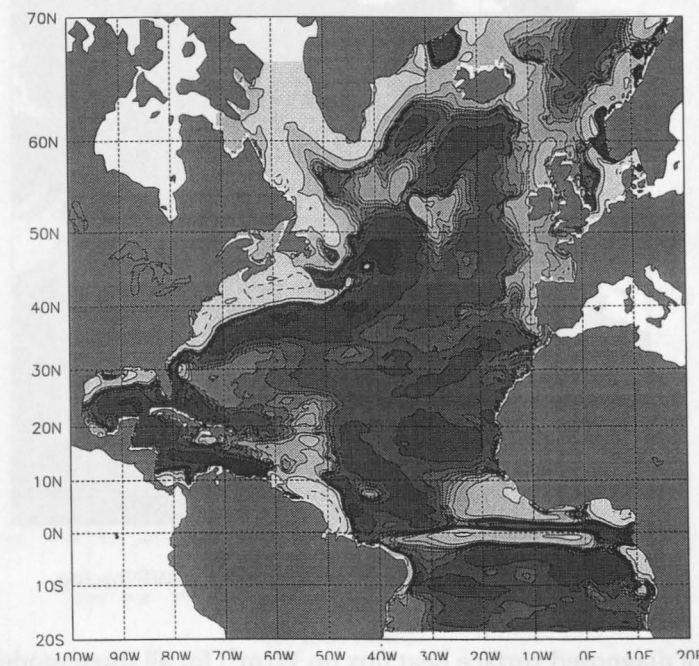


Figure 4.22: Maps of diagnosed surface heat flux (in W/m^2) for all three models, and for the ECMWF forcing.



(a) LEVEL



(b) ISOPYCNIC

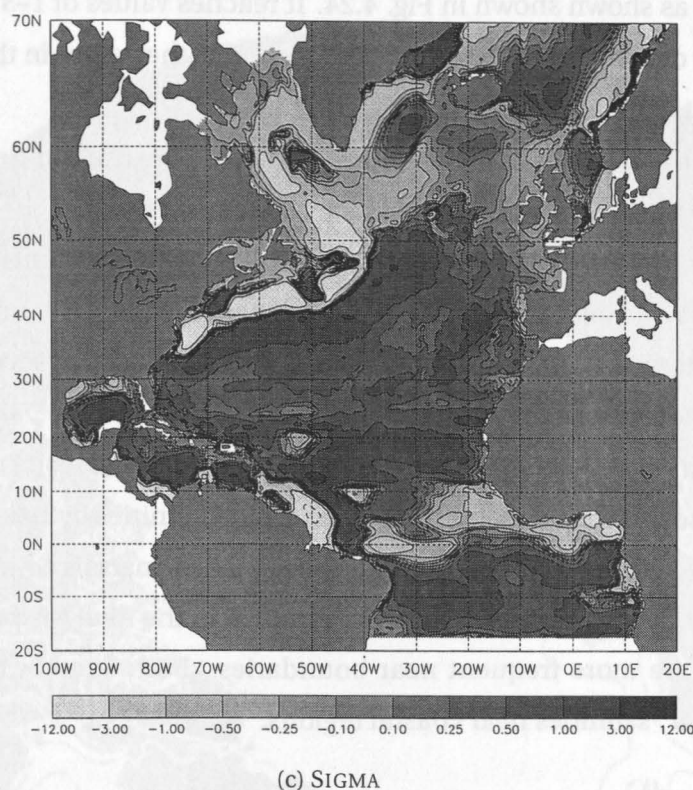


Figure 4.23: Maps for diagnosed surface freshwater flux (in m/y) for all three models.

The DYNAMO-integrations are no exception. The diagnosed net surface freshwater flux for the three models (Fig. 4.23) is qualitative fairly similar, however with considerable quantitative differences. Over wide regions, all models show net evaporation which is strongest in the Gulf Stream region and also between 10–20°N, reaching magnitudes of more than 1.5 m/y. Over most of the interior subtropical gyre the flux is weak, presumably for the same reasons that also lead to a weak heat flux here. Net precipitation occurs predominantly in high latitudes, and also in a small region near the Equator reflecting the imprint of the ITCZ on the sea surface salinity. The net fresh water flux into the ocean is particularly large in ISOPYCNIC where it exceeds 3 m/y around Greenland and over much of the Labrador Sea, and also on the northwestern flank of the Gulf Stream. The pattern in the other two models is similar, with somewhat smaller values.

Several effects can contribute to these high estimates. The principal reason for the apparent large net precipitation around Greenland is probably the lack of a sea ice component in all models. The annual mean net melting rate of sea ice which is transported through Denmark Strait and Davis Strait into the subpolar North Atlantic has been determined by HARDER

and LEMKE (1997), as shown in Fig. 4.24. It reaches values of 1–3 m/y in the Irminger Sea, and 0.5–1 m/y over much of the Labrador Sea, hence the values in the melting regions in Fig. 4.23 should not be interpreted as air-sea fluxes.

Deficiencies of the ocean model dynamics can of course also not be ruled out. Near the Gulf Stream large values in both directions can be expected when the path of the model's Gulf Stream differs from the observed pathway. The specific problems of ISOPYCNIC in the Labrador Sea may be a consequence of the mixed layer formulation, and it is plausible that the attempt to restore the surface salinity over a water column of several hundred meters leads to high surface fluxes.

Finally, the errors in the forcing functions may contribute. In fact it is not unlikely that the objective analysis scheme used to construct the surface salinities by LEVITUS (1982) may lead to a systematic bias near boundaries, due to i) the objective analysis which tends to smear out smaller scales near boundaries, and ii) due to removal in the analysis procedure of extremal values which also are more frequent near boundaries. Both sources of bias contribute to higher than “normal” salinities near coastal regions.

4.3.2 Mixed layer

The surface mixed-layer is the region where temperature and salinity are vertically homogenised through (wind-induced or convectively generated) turbulent transports. For that reason the mixed-layer structure has a significant effect on the surface heat flux, and in most simulations, including the present ones, also on the freshwater fluxes which depend on surface (i.e. mixed-layer) values. In the mixed-layer formulation of ISOPYCNIC, the velocity is vertically homogenised whereas the other models in principle permit shear within the mixed-layer, although the vertical friction of $10^{-3} \text{ m}^2/\text{s}$ can be expected to mix momentum downward rather efficiently.

The mixed-layer depth (MLD) in winter is shown in Fig. 4.25. By and large, the depth distributions in all models agree roughly with each other. At mid-latitudes, all models exhibit a characteristic pattern in ML-depths, with a gradient which is strongest in LEVEL and SIGMA, normal to a line from Florida to Cape Finisterre which separates depths shallower than 150 m to the south from greater depths to the north. All models show also a significant increase in ML depth in the northeastern basin and in the Irminger Sea, reflecting the heat loss that is experienced by water moving with the North Atlantic Current and with the general circulation in the subpolar gyre. Values exceeding 500 m are found in this region, in good agreement with the estimates by STAMMER et al. (1987) and MCCARTNEY and TALLEY (1982). Another sharp

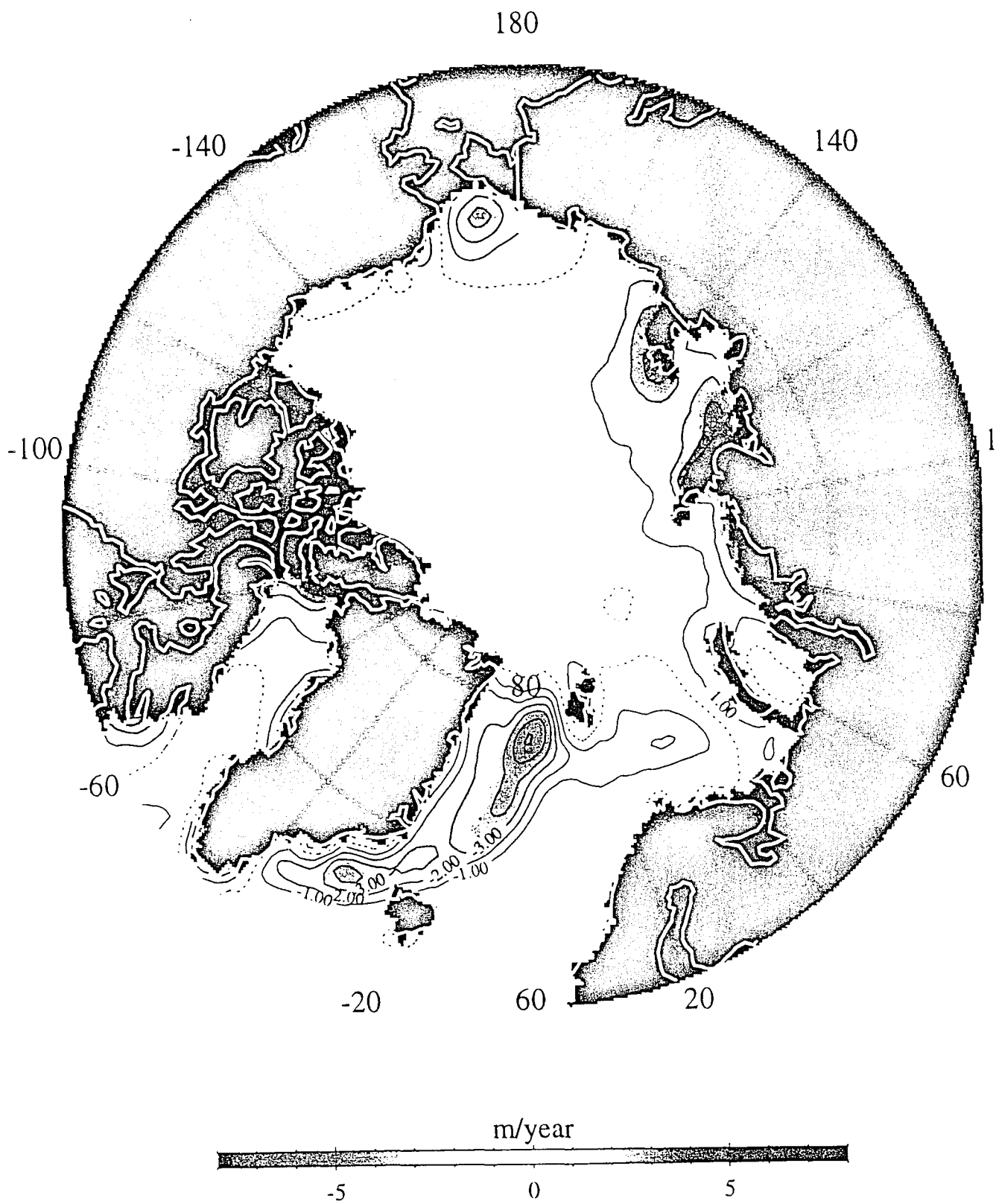
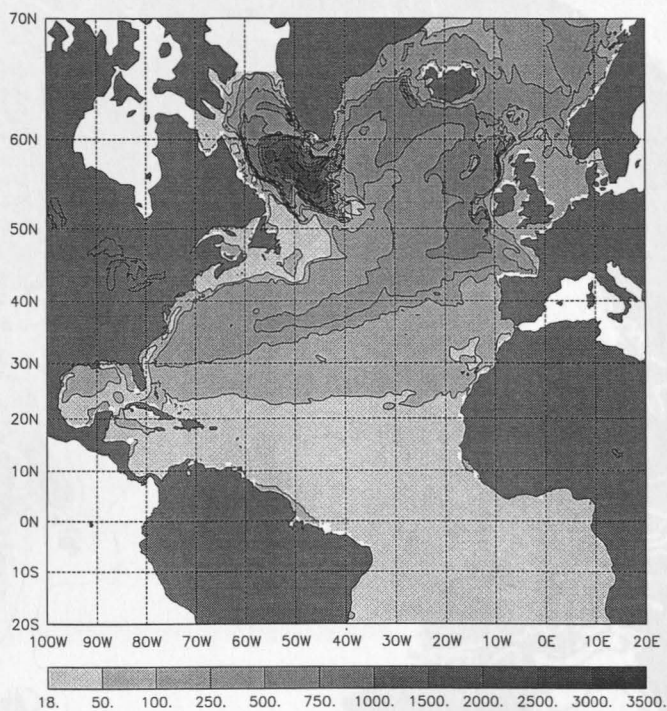
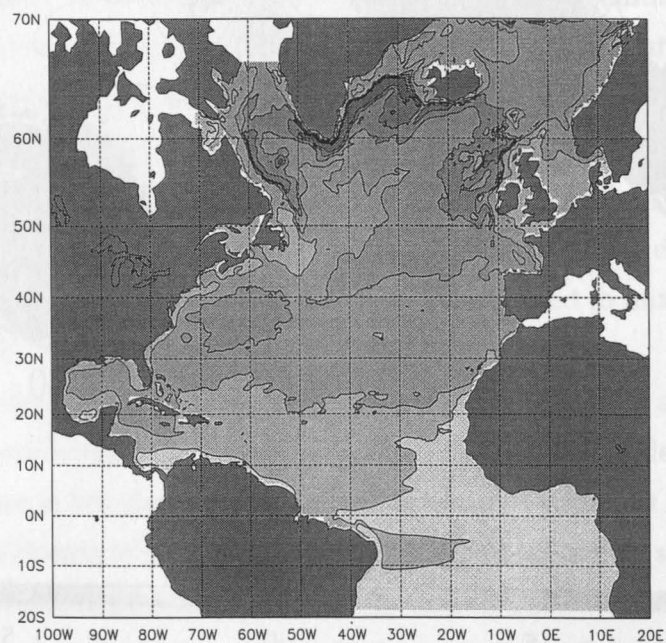


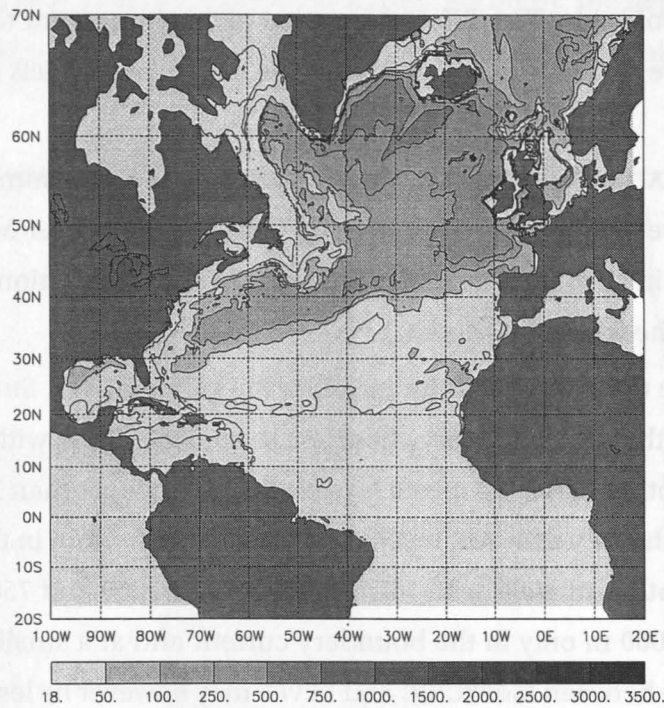
Figure 4.24: Average net freezing rate from 1986-92, in m/y.



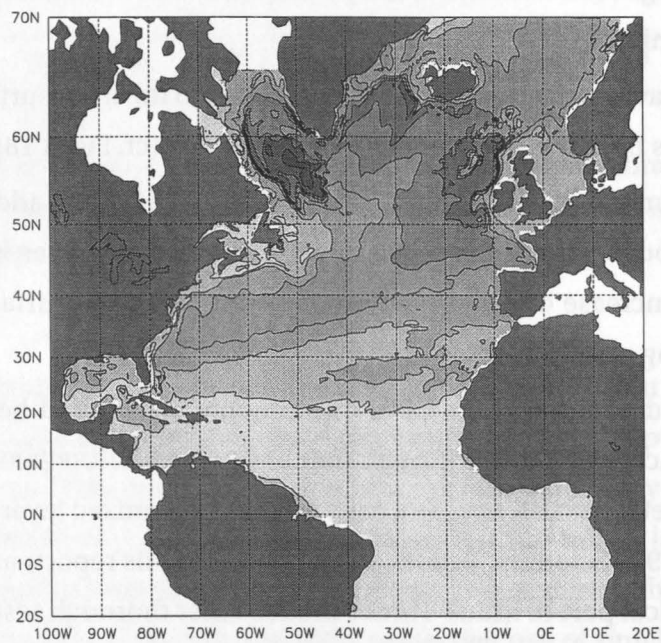
(a) LEVEL



(b) ISOPYCNIC



(c) SIGMA



(d) KTmix

Figure 4.25: Wintertime mixed layer depth for all three models, and for the KT mixed layer experiment of LEVEL.

gradient is found in the Gulf Stream region, with depths of 150–250 m on the warm side and 50 m or less on the cold side. The different location of this front point towards differences in Gulf Stream structure (see below). Over most of the tropics, all models predict ML depths of less than 50 m.

LEVEL and SIGMA have no explicit ML-algorithm in the intercomparison experiment, other than the convective mixing scheme. The fact that the winter ML depths are nevertheless fairly similar in all models indicates however that the convection is indeed the crucial factor determining the winter ML-depth.

The region where the estimates differ mostly is the Labrador Sea. SIGMA has an extremely shallow mixed layer that is nearly everywhere less than 250 m deep, without any sign of deep convection. On the other hand, the mixed layer in LEVEL is deeper than 1,000 m over much of the cyclonic gyre in the Labrador sea, reaching almost to the bottom in the centre. ISOPYCNIC is in between both other models, with ML depths between 250 and 750 m over most of the region, exceeding 1,000 m only in the boundary current and at a small spot near 54°W and 57°N. The difference between ISOPYCNIC and LEVEL may however be less dramatic when one considers that the ML depth is differently defined in both models; it is a prognostic parameter in ISOPYCNIC but diagnosed from the density stratification in LEVEL, and somewhat sensitive to details of the definition.

The unusual behaviour of SIGMA is obviously related to the near-surface circulation which advects warm waters from the south into the Labrador Sea (cf. Fig. 4.18(c)).

All LEVEL experiments except KTmix were conducted without additional Kraus-Turner type mixed layer model. The depth of the mixed layer in these cases is diagnosed from the density field by an increase of 0.01 sigma units compared to the surface, a criterion widely used in the modelling community.

A recent, detailed analysis of the CME experiments, which had been performed with a mixed layer model according to the formulation of CAMP and ELSBERRY (1978), led to uncertainties of the energetic consistency. As a consequence, the mixed layer code was completely rewritten (DENG, 1995 Scientific Report; Introduction in this report) and now correctly represents the mechanical part of Kraus-Turner model. After thorough testing in the coarse resolution level model (J. DENG, pers. comm., 1996), LEVEL was set up with exactly the same mixed layer parameters as ISOPYCNIC. Figure 4.25(d) depicts the mean wintertime mixed layer depth for this experiment. Values are generally closer to ISOPYCNIC than in the intercomparison experiment. It is at first remarkable that additional, wind-induced stirring at the surface shallows the mixed layer in the central Labrador basin. Whereas in the intercom-

parison experiment convection reached the bottom in the central Labrador Sea, somewhat more realistic values of 2500 m are reached in KTmix. Secondly, the depth of the mixed layer in the Irminger Basin is increased up to 1500 m, very similar to the mixed layer structure of ISOPYCNIC.

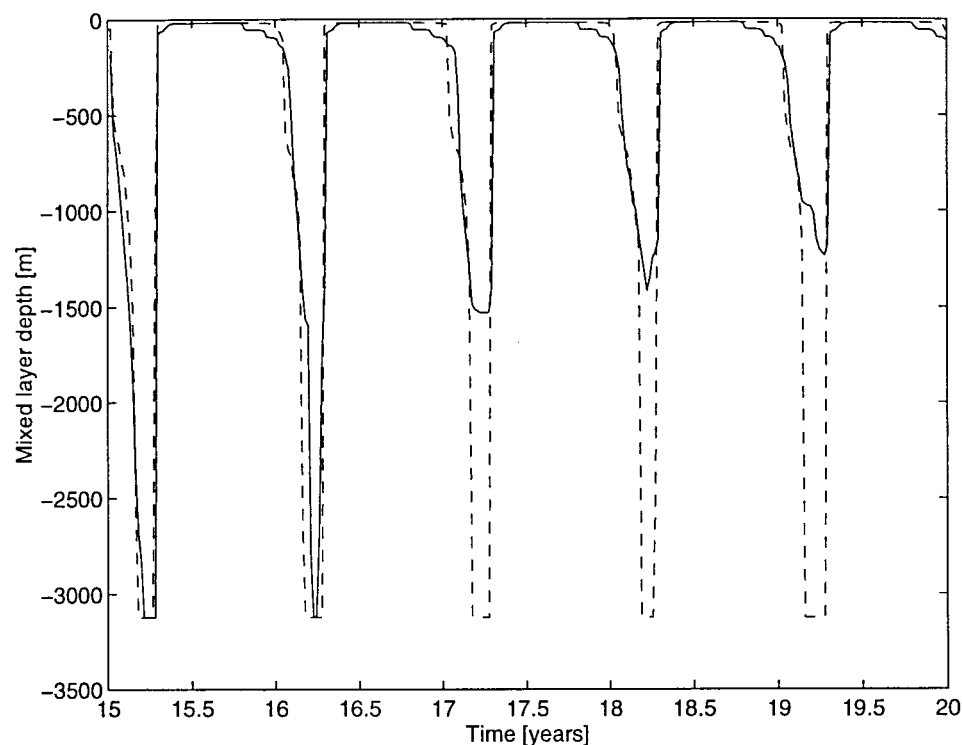


Figure 4.26: Temporal evolution of mixed layer depth in the central Labrador Sea (57°N , 54°W). Dashed: LEVEL intercomparison experiment. Solid: LEVEL KTmix experiment with alternate mixed layer formulation.

The time evolution of the mixed layer depth at a single station in the Labrador Sea (Fig. 4.26) clearly shows a shallowing of the wintertime mixed layer depth, resulting in a maximum of 1200–1500 m. This is the depth to where the integral energy balance is evaluated. From former CME experiments, it may be expected that the mixed layer depth will never exceed this level in further integration of the model. This is due to the calling sequence to subroutines of the numerical model, and their respective use of timelevels, and may eventually be overcome by a change of the calling sequence, choice of another algorithm for the parameterisation of deep convection or a complete redesign of the present memory layout of the model.

Increasing interest in mixed layer dynamics arose recently from coupled physical–bio-

/geochemical numerical modelling. Several ongoing studies are based on simplified biological models according to FASHAM et al. (1990), coupled to basin-scale (OSCHLIES, 1997, pers. comm.) or regional (DETERMANN, 1997, pers. comm.) physical models and are of relevance to the European community with respect to fisheries. A proper representation of mixed layer dynamics plays a crucial role for the nutrient budgets. As an example, the temporal evolution

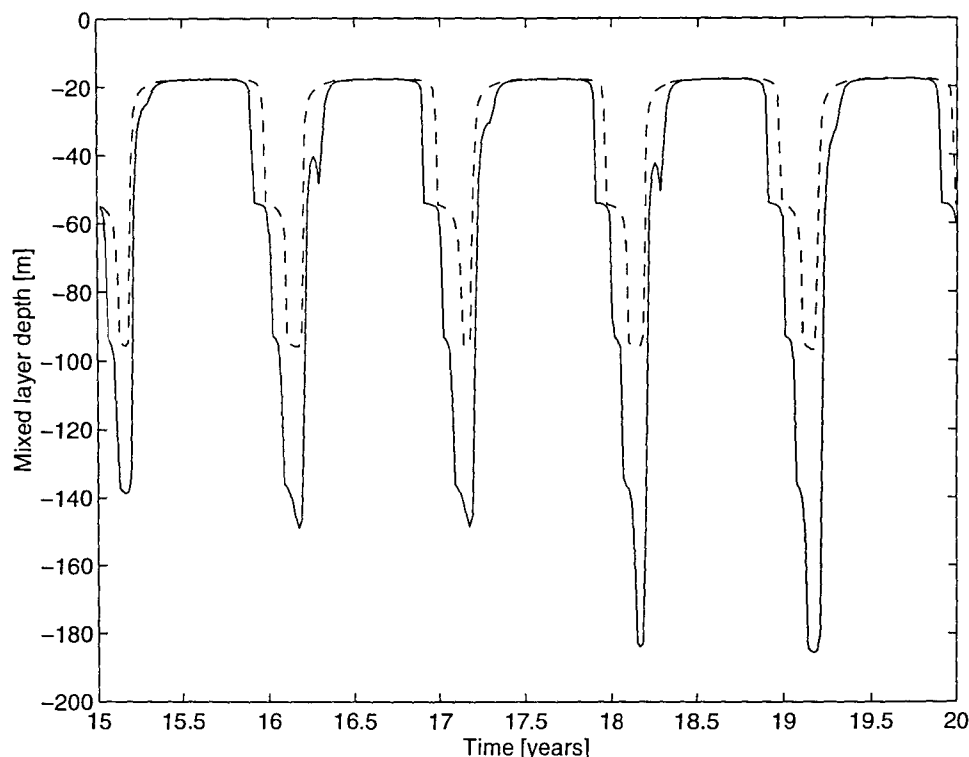


Figure 4.27: Temporal evolution of mixed layer depth in the North East Atlantic (40°N , 15°W). Dashed: LEVEL intercomparison experiment. Solid: LEVEL KTmix experiment with alternate mixed layer formulation.

of the mixed layer depth in the central Iberian Basin, off Lisbon, is depicted in figure 4.27. Whereas in the intercomparison experiment without additional mixing due to the windwork, the wintertime mixed layer depth never exceeded 90 m depth, with the Kraus-Turner type mixed layer model, driven by the climatological monthly mean atmospheric friction velocity, a significant deepening of the wintertime mixed layer depth may be observed. Strong winds in late fall lead to an earlier mixed layer deepening, and a later shallowing is revealed in late March. Overall, the Kraus-Turner type mixed layer model leads to a more realistic seasonal cycle.